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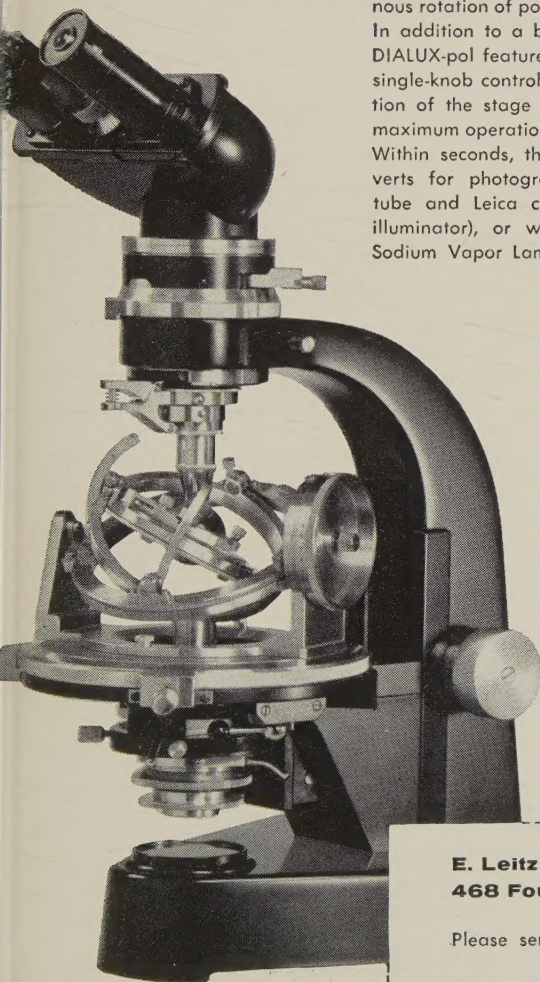
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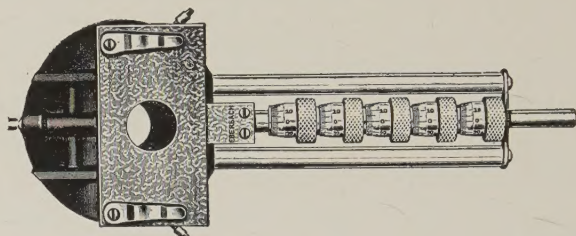
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The above statement is part of a letter sent out last November to about 200 members by a volunteer group consisting of 20 Fellows. They set out to form a "*Fifty-plus Committee*." Each member was to pledge *at least* \$10.00 per year for five years. If 100 members joined this would mean at least \$5,000 and quite possibly double that amount. Through the efforts of this committee, the total pledged is already approximately \$10,000 from about 100 members.

It is obvious that a strong feeling exists among our membership that this is a timely campaign, and it seems that a general appeal should result in at least 250 members. If any member wishes to join, please address a note to our Treasurer, Miss Marjorie Hooker, stating that you pledge \$10.00 (or more) per year for five years. (Several have paid not less than \$50 as a single pledge. Already about \$2,000 in cash has been received.)

It would seem that with a little more effort, we could reach a total of \$25,000. This is well within the potential of our membership, and would serve as a great stimulus to go out after a much larger endowment fund by the time of our golden anniversary in 1969. Such an endowment would give us real economic stability. We could plan and carry out ideas for the advancement of the mineralogical sciences hardly dreamed of at present. Colonel Roebling's faith in us would be more than realized.

If you would like to have a part in this effort, please send your pledge without delay. We hope to print the names of the members of the committee in the next issue of our Journal.

THE AMERICAN MINERALOGIST

JOURNAL OF THE MINERALOGICAL SOCIETY OF AMERICA

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Nos. 3 and 4

SOME TEXTURAL FEATURES OF MAGMATIC AND METASOMATIC ROCKS*

G. E. GOODSPEED, *University of Washington, Seattle 5, Washington.*

ABSTRACT

The textural features of magmatic rocks have been described with the emphasis placed upon their relationship to the crystallization sequence of the magma. Examples have been chosen from the basaltic and diabasic dikes in the vicinity of Cornucopia, Oregon, the shonkinite-granite porphyry sequence at Yogo Peak in the Little Belt Mountains of Central Montana, and a few other localities. The textural differences of the orthomagmatic as compared to the late magmatic or deuteric pattern of crystallization have been noted. Where it has appeared that the deuteric fraction had been involved in mass flowage the term deuteromagmatic is suggested.

For metasomatic rocks, textural features to illustrate various stages of crystal growth have been described, such as the development of porphyroblasts, glomeroblastic aggregates, and the final crystalloblastic pattern. Most of the examples are taken from occurrences near Cornucopia, Oregon, and Buffalo Hump, Idaho, with a few references to other localities.

The textural features resulting from the mobilization of metasomatized rocks, such as rheomorphic dikes and rheomorphic breccias have been briefly noted. These commonly show magmatic textures superposed on metasomatic ones. For many igneous rocks, however, later crystalloblastic textures are superposed on the earlier orthomagmatic textures. Textural features alone are not always adequate for petrogenetic interpretations but for some occurrences like granitic intrusions they may furnish clues as to whether the body has evolved from an orthomagma or from mobilized metasomatized material, namely a neomagma.

INTRODUCTION

The title of this address might suggest that textural features alone could form the basis for distinguishing a rock of magmatic descent from one of metasomatic lineage. Such an implication is not intended since many other features contribute toward adequate genetic interpretations. Chief among these is that of the geologic setting, including the more immediate field relations of the rock mass. Indeed the petrographic study of rocks in thin sections, and especially in large thin sections, may be

* Address of the retiring President of the Mineralogical Society of America at the 39th meeting of the Society at St. Louis Missouri, November 7, 1958.

likened to a field study of a much contracted area but one which nevertheless may furnish many data of genetic importance.

Textural features of igneous or metamorphic rocks, as seen under the petrographic microscope, may be thought of as a pictorial record of the struggles of individual minerals to adapt themselves to the changes of their environment. Some textural patterns may resemble a naturalistic painting or design and these usually afford the basis for a ready explanation of origin. Other textural patterns, however, are more like modern art, and their interpretation from a genetic standpoint is influenced by the experience and imaginative ability of the viewer.

Many years ago Holmes (1921) stated: "Every texture and structure is a hallmark stamped on a rock by some process through which it has passed, and one of the most important objects of petrological study is the correlation of individual textures and structures with processes, and of the combinations of textures and structures with the succession of processes in time." More recently Barth (1952), in reference to metamorphic rocks, made this comment: "A great many textures can be interpreted in a variety of ways, depending upon the dominant belief of the observer." In a recent paper on the regional metamorphism and granitization in the Central Pyrenees, Zwart (1958) quite rightly points out that textural features are second order criteria. He states: "The occurrence of a so-called igneous texture versus a crystalloblastic texture says, of course, nothing directly about the mode of emplacement of the granite. Together with many other observations it may be important, but it will never have such a conclusive value as observations which are concerned with the emplacement itself."

Two contrasting environmental conditions must be considered; first, the growth of crystals from a liquid silicate melt or from a pasty, but mobile magma, and second, the development of minerals by metamorphic processes in a solid rock.

As a magma cools and crystallizes, the earlier minerals are formed in an environment conducive to the development of well-formed crystals, then as the initial fluidity dwindles, the later minerals are restricted by the earlier ones which are also subjected to magmatic reactions. In the final stages of consolidation minerals may form in a nearly solid medium. Many small igneous rock bodies such as diabase dikes display textural features which clearly indicate varying rates of cooling as well as the intersertal or intergranular textures which are good records of the magmatic stage of crystallization. Some of these occurrences also show textural features resulting from the crystallization of the late magmatic stage or of the deuteric fraction, and these commonly resemble the crystalloblastic textures of metasomatic rocks.

In metasomatic rocks, however, minerals are formed in an essentially solid medium by reactive penetrating solutions or emanations under conditions of rising temperature. This means that a growing crystal develops by pushing aside, surrounding, or replacing the original minerals of the metamorphic rock or by a combination of these processes. Hence, the initial crystalloblastic textures displayed by these growing crystals are quite different from those textural features resulting from the early stages of magmatic crystallization.

With continued increase in temperature, some metamorphic rocks may become mobile and their final stages of crystallization will be quite similar to that of some rocks formed from a fluid magma, so that their textural features will look like those of rocks of orthomagmatic descent. This convergence of textural features in rocks of diverse modes of origin means that it might be most difficult if not impossible to make adequate genetic interpretations based on textural features alone. Such is the case, unless the preservation of earlier formed minerals or relict textural features can furnish some clues pertinent to the genetic history of the rock. In other words, do the earlier formed minerals or relict textural features indicate crystallization from a magma or do they suggest an earlier period of metamorphism? The question as to whether or not the earlier minerals represent an earlier magmatic or metamorphic stage is further complicated by the fact that many minerals can be precipitated from a magma by a decrease in temperature or can be formed during metamorphism by increasing temperature. Mineral associations and their textural features may aid in such interpretations.

At present, textural features are largely descriptive, and genetic interpretations are much influenced by assumptions as to whether a particular rock is of magmatic or metasomatic origin. Neomagmatic rocks formed by rheomorphism have intrusive field relations and are properly classed as igneous, although their line of descent is through metasomatism rather than from an orthomagma.

In the future it is hoped that many textures of rock-forming minerals will be produced synthetically under known and controlled conditions of temperature and pressure, thus affording more accurate data for genetic interpretations.

TEXTURAL FEATURES OF THE TERTIARY BASALTIC AND DIABASIC DIKES AT CORNUCOPIA, OREGON

General field relations of the Tertiary dikes

One of the most conspicuous geologic features of the Wallowa Mountains of Northeastern Oregon are the numerous Tertiary basaltic and

diabasic dikes which transect the older sedimentary, metamorphic and granitic rocks. In the vicinity of the old mining town of Cornucopia, which is in the southeastern portion of the Wallowas, the dikes stand out as dark brownish bands in vivid contrast to the light-colored granites.

In discussing the basalt dikes of the Blue Mountains of Oregon, Waldemar Lindgren wrote, "But in no place are they exposed on such a magnificent scale as in the Bonanza Basin near Cornucopia" (W. Lindgren, 1901). Not only are the dikes well exposed on the steep walls of glacial valleys and cirques, such as the Bonanza Basin, in this rugged area of nearly 5000 feet of relief, but they have been revealed in many of the underground mine workings including several long adits having a total length of over 18,000 feet. An area of about two square miles in the vicinity of Cornucopia is crossed by over forty basalt and diabase dikes with a prevailing northerly or northeasterly trend. At lower elevations most of the dikes have steep dips and transect the conjugate joints. At higher elevations many of the dikes follow these westerly or easterly dipping joints. The dikes range in width from 25 to 100 feet and usually have parallel walls, although local widening and changes in strike and dip are characteristic where they have intruded planes of weakness, such as some of the vein shear zones. The relation of the dikes to the veins were very accurately mapped by H. F. Anderson and J. P. Dunn in connection with their work for the Cornucopia Gold Mines Company and have been described by Fredrickson (1950).

It is apparent that there was more than one episode of dike injection; earlier diabasic dikes are transected by basaltic dikes which in turn are cut by later diabase dikes and then by basalt dikes. The fine-grained basaltic dikes commonly exhibit well defined columnar structure and are more resistant to weathering than the coarser grained diabasic dikes. Many of the dikes have narrow glassy selvages, usually only a fraction of an inch in width, others do not have these selvages but are finely crystalline at the actual contact with the wall rocks. The finer-grained borders of some of the dikes are, locally, two or three feet in width. There is also a marked contrast in grain size between the dense basalt dikes and the coarse-grained diabase dikes. The main mechanism of intrusion was one of dilation, and where the walls are parallel, earlier rock units and dikes transected at acute angles display appropriate offsets.

A few miles north of Cornucopia in a 200 square mile area of the Northern Wallowa Mountains, H. W. Smedes (1959) has recently mapped about 160 Tertiary basaltic and diabasic dikes, over 30 of which range from one to three miles in length with one nearly four miles long. Here, too, the dikes have a northerly trend, some northeasterly or north-westerly, and only four with an east-west strike. Almost all of these dikes are nearly vertical or have steep dips.

This display of Tertiary volcanism leaves little doubt that the dikes were originally feeders for some of the near-by extensive lava flows of the Columbia River Basalts. In an area two miles northeast of Cornucopia, H. F. Anderson mapped 22 flows averaging 40 to 50 feet thick forming a ridge having an elevation of about 7100 feet. He also noted 10 more flows on a higher ridge $\frac{1}{4}$ mile to the north. Although in the Cornucopia area no distinct exposures have been found to show the actual merging of a dike into a flow, such a one was described by Fuller (1927) in western Idaho on the canyon walls of Rock Creek, a tributary of the Salmon River. Here, some seventy miles northeast of Cornucopia, there is a remarkably good outcrop of a multiple dike which definitely extends upward and laterally into flows.

The general structural pattern of these Tertiary dikes as well as their association with extensive basaltic flows leaves little doubt that they were caused by fracturing so profound as to permit the upwelling of locally liquid sima. The fine-grained borders of the dikes probably represent the rapid chilling of the initial upward surge of magma and also served as insulators to retard heat losses as the dike fissures became conduits for the free-flowing magma. The coarse-grained central portions of the dikes represent the relatively slower cooling of the final phase of the magma.

With very few exceptions, the Cornucopia dikes are quite free from inclusions so that their crystallization was not complicated by the effects of contamination and assimilation. These effects, however, are very noticeable in one dike which is filled with inclusions for about 100 feet back from its wedge-shaped lateral termination. Some of the dikes do show deuteric features and a few contain later veinlets with low temperature minerals. None of the dikes have suffered from extraneous metamorphism.

Textural features of the magmatic stage of crystallization

The relatively simple mode of occurrence of these Tertiary dikes as well as their nearly uniform composition means that the textural features can furnish some clues, not only with regard to the initial physical state of the eruptive magmas, but also with respect to their crystallization sequence which in most of the dikes was uninterrupted by any addition of foreign material. The data for these textural features was obtained by the study of thin sections from several hundred specimens taken from surface and underground exposures and diamond drill cores.

In his Descriptive Petrography of the Igneous Rocks, Johansson (1939) makes this statement with regard to basalt and diabase: "In the United States it is customary to call dikes or intruded sheets of basaltic composition 'diabase' since they usually have ophitic texture, but there

is no reason why they should not be called basalt dikes, if desired, since the mineral composition is exactly the same, and there may be no difference in texture." Yet one of the most noticeable textural features of the Cornucopia dikes is the difference in grain size between the dense basaltic dikes and the coarse-grained diabase ones. For example, a thin section of a specimen from near the center of a dense basalt dike 30 feet wide discloses a mat-like aggregate of labradorite microlites (0.2 to 0.08 mm.) with numerous interstitial clouded grains of mafics (0.05 mm.) and magnetite with about three per cent of brown glass. An area of one square millimeter contains 490 mineral grains: 160 plagioclase microlites, 230 grains of mafics and 100 grains of magnetite. In contrast to this high concentration of centers of crystallization in basalt, a thin section from the central part of a coarse-grained diabase dike, 20 feet wide, shows on an average of 15 crystals or parts of crystals per square millimeter: i.e. 4 labradorite, 5 mafics, 6 magnetite.

Several years ago, H. F. Anderson (personal communication), who was at that time the resident geologist of the Cornucopia Gold Mines, came to the conclusion that the state of the magma in the underlying reservoir prior to injection was a determinative factor in reference to the final textural features of the basalt and diabase dikes. He pointed out that at Cornucopia some large dikes have a basaltic texture and others a coarse diabasic texture and also that some of the nearby Tertiary lava flows are fine-grained basalts and others coarse-grained diabbases (Fig. 1).

As stated by Turner and Verhoogen (1951, p. 43), small crystal grains have a higher chemical potential than larger ones, so that for a slight degree of undercooling only large nuclei are stable. Hence, in a greatly undercooled magma the resulting rock would show many crystal grains representing crystallization centers which were unable to form larger individuals because of a rapid increase in viscosity. Under conditions of a slight degree of undercooling larger crystals will form at the expense of small ones.

Since it is probable that the Tertiary basaltic and diabase dikes of this region were successively intruded over a considerable length of time, it is reasonable to assume that there may have been some variation in the temperature of the source pockets of magma in the liquefied sima. Moreover this might also account for some slight differences in chemical composition of the successive dike intrusions.

Recently (1958) J. F. Lovering in a paper on the nature of the Mohorovicic discontinuity has suggested that "The eclogitic material which originally existed between the original and final levels of the discontinuity will transform to basaltic material with a volume increase of something like 15 per cent. It is the volume increase which then provides the mech-

anisms for raising the overlying crustal column relative to adjacent regions under which the Mohorovicic discontinuity has been unaffected." Under this hypothesis variations in composition of the eclogitic material would be reflected in the composition of the basaltic material which, also due to the 15 per cent increase in volume, would have a greater irruptive potentiality.

The fine-grained borders of the dikes represent the rapidly cooled initial magma, and they commonly contain a few euhedral crystals of labradorite and pyroxene in a glassy or very fine-grained groundmass. Most

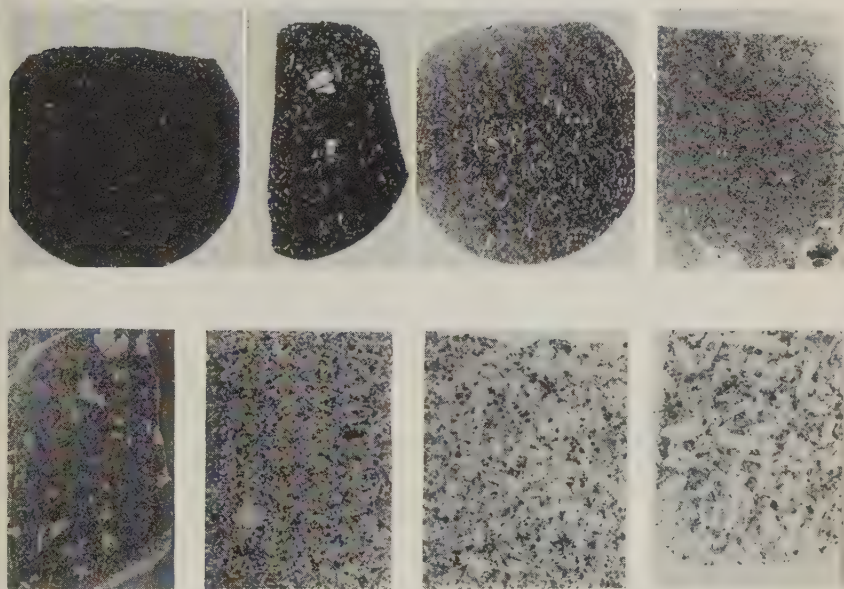


FIG. 1. Photographs (ordinary light) of four thin sections from diamond drill cores of basalt dikes (above) and of four thin sections from specimens of Cornucopia diabase dikes (below).

of the plagioclases are clear, twinned, and relatively free from fractures. Flow alignment shows that these crystals were originally free swimming in mobile liquid environment. Some narrow (3 to 4 inches) dike apophyses show skeleton crystals of the dominant minerals.

The textural features displayed in thin sections of specimens from the central parts of many of the dikes are the record of the consolidation of the final surge of magma at the cessation of the fissure eruption for which a particular dike was the feeder. The magma at that time may have been partially crystallized with numerous relatively large individual crystals; or it may have had, as in the case of the basaltic dikes, a great many very

small crystals. The coarser-grained diabase dikes have intersertal, subophitic, or intergranular textures commonly modified by porphyritic textures.

Many of the porphyritic varieties show glomerophyric aggregates of large tabular labradorite crystals ranging from rather loosely grouped ones to very compact groups which approach the form of a single crystal. This suggests the possibility that feldspar crystals during a mobile stage of the magma preceding consolidation collided with one another and that some of them became stuck together in a haphazard arrangement. The material included in the triangular interstitial spaces of these aggregates depends upon the state of crystallization of the magma at the time and place of their formation.

Rectangular inclusions of glass or earlier pyrogenetic minerals are noticeable in many phenocrysts; and, in general, these inclusions are arranged parallel to the 010 faces of the labradorite phenocrysts. It seems quite probable that some of these crystals have been built up by the attachment and welding of smaller crystals. Where the weld is perfect, no material is included; if, however, two or more smaller crystals adhere to the tabular surface of the 010 face but fail to grow together, they will be separated by a rectangular re-entrant filled with the mesostasis (Fig. 2). Then if other crystals grow across this gap, the mesostasis is completely enclosed and becomes an inclusion which represents the state of the mesostasis at the time and place of the formation of the phenocrysts. In other words, a phenocryst of intratelluric origin would include material quite different from the mesostasis surrounding the phenocryst. Inclusions of glass are quite common; some consist of glass and pyrogenetic minerals or of pyrogenetic minerals alone. In some cases, as will be noted later, what look like inclusions may be the result of magmatic replacement.

The preference for growth along the 010 face of labradorite appears to be contingent upon its tabular habit when crystallized from a magma. Buerger (1947) in his paper on the relative importance of crystal faces states: "When a molecule lands on the crystal surface so as to continue the crystal structure, then the energy of the bond between the molecule and the crystal is (to first approximation) proportional to the area of the surface joining the molecule to the crystal. If several sites on the crystal are available, the preferred site (neglecting differential thermal effects) is the one with maximum bond strength. With the simplifying assumptions of this section, this is the site which offers the greatest area of attachment between crystal and molecule."

It should be pointed out that by no means all labradorite phenocrysts have a mode of growth as suggested above. Many, no doubt, develop as

single crystals. Niggli (Parker translation, 1954) mentioned the effect that seed crystals have in promoting crystallization. He wrote: "It is a general rule that the formation of nuclei settling on a solid surface of any kind requires less work than does the formation of freely suspended nuclei" (p. 469).

Although most of the feldspars in the Tertiary dikes display sharp outlines, a few of them in some of the diabase dikes have ragged borders



FIG. 2. Photomicrograph (plane light) of labradorite phenocrysts in basalt, Cornucopia, Oregon. Note: on the marked 1.7 mm. crystal the two smaller (0.2 mm.) attached crystals.

in contact with pyroxene. These outlines are apparently caused by magmatic corrosion. In some sections the paragenesis is clearly shown by the cusp-like borders of the mafics against the feldspars and also by the apparent penetration of mafics along the albite twinning planes. Where the apices of these cusps happen to be truncated by the plane of the thin section, the plagioclase looks as if it had included numerous small rounded mafic crystals; whereas, in reality, the final crystallization of the mafics is later than that of the feldspars. Some of these corroded plagioclases show thin delicate relics which extend into the mafic-rich groundmass, thus indicating that, at this stage of crystallization, the magma was in a rather static non-mobile state.

Another feature of most of the larger plagioclases is the presence of

numerous irregular fractures which in some phenocrysts have an irregular veined-network pattern (Fig. 3). In some glomerophyric groups the fractures cut across the component feldspars and also penetrate glass and mafic mineral inclusions. Johannsen (1939) mentions fractures in plagioclase in basalt and states: "Cleavage on (010) and (001) often shows, and there may be cross-fractures along no special face, as well." In a more recent paper C. A. Zapffe, C. O. Worden, and Carl Zapffe

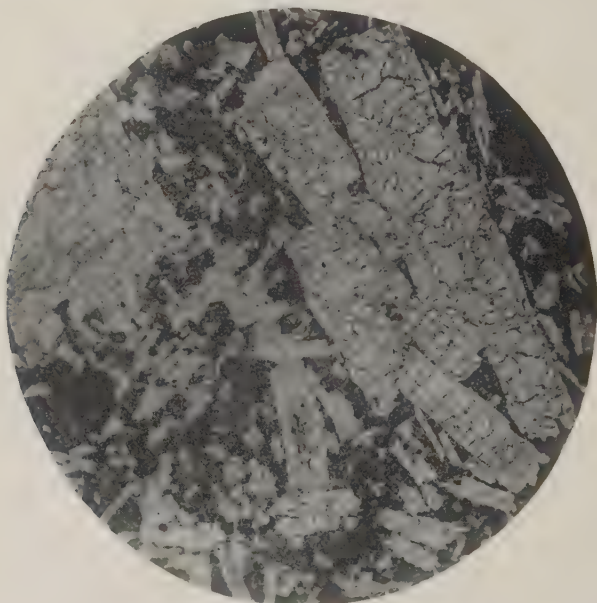


FIG. 3. Photomicrograph (ordinary light) of part of a thin section of the matrix of the contaminated dike. Note the inclusions in the glomerophyric aggregate of labradorite and the irregular fractures. $\times 30$.

(1951) have shown many interesting fractographs of minerals including the feldspars. Most of the small plagioclases, 0.1 to 0.2 mm. are devoid of fractures although some of them are slightly fractured. These fractures play an important role in the final stages of the crystallization history in that they provide minute channelways for late magmatic and deuteric alteration. In a few of the plagioclases the fractures form a pattern of closely spaced parallel lines and are perhaps similar to what has been described as lineage structure.

Several possibilities may be suggested to explain the origin of the fractures in the plagioclases:

- (1) The crystal mesh of the cooling magma may have been subjected

to a uniform internal stress which was resisted by the criss-cross pattern of the feldspars characteristic of the diabases. A possible analogy is the decussate structure found in some metamorphic rocks.

(2) Crystallization of the pyroxenes pushing apart as well as partially replacing some of the earlier plagioclases may also have contributed to the internal stress of the almost completely crystallized magma.

(3) Some of the larger feldspars may have been of intratelluric origin and thus formed under conditions of considerable pressure. The sudden diminution of this pressure as the crystals were transported upward in the dikes to where a much lower pressure prevailed might cause an expansion in the crystals thus producing fractures.

(4) Some of the fractures may be similar to lineage structure. Based on evidence derived from artificial crystals, several interpretations regarding the origin of lineages have been advanced. Tuttle and Twenhofel (1946) suggest that crystals grown at higher temperatures, and hence under conditions of greater agitation and energy, have fewer structural defects than those grown at lower temperatures.

(5) Additional internal stress might be produced at the initial stage of the development of a deuteritic fraction with its complement of less dense and hydrous minerals. At a slightly later stage deuteritic products permeate the previously formed fractures in the plagioclase crystals and may cause more fracturing (Fig. 3).

The possibilities just listed are by no means mutually exclusive, but petrographic data are not sufficiently definitive to draw conclusions as to which play dominant roles. Future work on the textural features developed under known conditions from artificial melts would be of much aid in clarifying this problem.

Next to the plagioclases the most abundant minerals of these Tertiary dikes are pyroxenes. Of these augite and pigeonite are the most common varieties, with a few representatives of the enstatite-diopside series. Euhedral pyroxenes are characteristic of the finer-grained border zones of the dikes. In the central parts they occur either as interstitial anhedral grains or as larger anhedral individuals surrounding the net of earlier-formed feldspars.

The next most abundant mineral is magnetite which occurs in small discrete grains and also more conspicuously in the form of large irregular shaped grains molded around the plagioclases and other earlier formed minerals. Therefore, magnetite can be interpreted both as early magmatic and late magmatic in origin.

Several other minerals are present in minor amounts in a few of the dikes. Olivine occurs in its usual rounded shape and commonly shows some alteration along cleavage cracks. Brown hornblende, of rare occur-

rence, forms discontinuous zones around pyroxene and partially replaces pyroxene where it follows the original outline of this earlier formed mineral. Relict islands of pyroxene are noticeable in some of the hornblende. The hornblende is embayed by late magmatic magnetite. In a few sections incipient skeletal feldspars can be seen in pockets of glass-residium. Perhaps some of the very few quartz grains in some of the dikes can also be considered to be late magmatic since they exhibit sharp boundaries and appear to have crystallized before the rock became a solid crystalline aggregate.

Textural features of the deuteric products

In addition to the minerals of the magmatic sequence most of the Tertiary dikes contain paulopost alteration products which usually occur interstitially or as replacements of the earlier formed pyrogenetic minerals or the glass residuum. A few of the basalt dikes appear to be free from alteration but some contain about five per cent and the diabase dikes usually have about 10 per cent of deuteric alteration products. These products have a wide variation in composition; greenish chloritic or brownish chlorophaetic material are the most common; micropegmatitic and myrmekitic intergrowths are conspicuous in some of the diabases, as well as tremolite, zeolites, carbonate, hydromica, and rarely a reddish micaceous mineral.

The intergrowths also vary considerably in occurrence; some are associated with interstitial quartz grains and some appear to have replaced an originally glass residuum. As a rule the finer myrmekitic intergrowths have replaced some of the plagioclases, either partially or wholly, with the intergrowth following but preserving the forms of the earlier feldspars. These relict outlines indicate that the rock was essentially solid at the time of the myrmekitic replacement.

In several dikes the deuteric fraction is noticeable in the field in the form of narrow irregular light-colored veinlets which are usually nearly parallel to or coincide with the columnar jointing. The material in these veinlets is quite similar to the deuteric products which occur interstitially, although it usually has a slightly different textural pattern.

A 2×2 thin section taken from a specimen from the central part of a diabase dike 30 feet wide, exposed in the lower adit of the mine, shows one of these narrow (1 mm. in width) veinlets. The groundmass of this veinlet has a slightly pinkish tinge and is so fine-grained that under low power magnification it appears to be nearly isotopic. Under high power, however, it is seen to consist of an aggregate of minute grains of quartz, untwinned feldspar and a considerable amount of fine dust-like ma-

terial which appears to be chiefly brown biotite. In this very fine grained groundmass are some relatively larger grains of quartz and feldspar and patches of micropegmatite. Some of these feldspars are albite and a few are orthoclase. Many of them are filled with minute inclusions of the very fine dust-like material of the groundmass. Some are quite anhedral, having amoeboid outlines with crenulated borders extending into the fine grained groundmass. A few of the albite crystals, however, are quite clear with sharp borders and have rectangular U-shaped outlines. The veinlet also contains a few crystals or fragments of labradorite crystals from the diabase. The pyrogenetic labradorite is commonly rimmed with albite and this also holds for those parts of the labradorite crystals which extend into the veinlet from the adjacent diabase.

Another noteworthy feature of this veinlet is the presence of a considerable amount of pinkish-brown mica which occurs in very small flakes with some suggestion of crystal faces. There are, however, several euhedral crystals of biotite, and some of these are included by the clear feldspars and by the albitic rims of the labradorite. Within the veinlet irregular patches of micropegmatite have formed late in the crystallization sequence since they not only surround the other minerals but also partially or wholly replace some of the late-forming feldspars as can be seen by the survival of the former outline of these crystals. Minor amounts of magnetite, chloritic material and carbonate are also present in the veinlet.

The textural features and the mineral composition of this veinlet are similar to those of some of the deuteritic products of the diabasic magma. It consists of that part of the deuteritic fraction which instead of remaining in the crystal mesh of the diabase become mobile enough to flow into a fracture, thus forming the veinlet. The crystallization sequence of the deuteritic material in the veinlet seems to have ranged from a stage of initial mobility to that of crystallization in a nearly solid medium. Since this material apparently behaved like a magma, although clearly deuteritic in origin, it is suggested that the term *deuteromagmatic* might be more appropriate than either magmatic or deuteritic. The chloritic material and carbonate in this veinlet may be explained as local hydrothermal products formed during the final stage in the crystallization sequence of the deuteromagma.

Other minute veinlets consist of very low temperature minerals such as zeolites, carbonate, and chalcedonic quartz, and exhibit textural features indicative of hydrothermal deposition. They may have been formed during the closing stage of magmatic or deuteritic crystallization. Another feature which may be accounted for by hydrothermal action are the

minute vermiform quartz veinlets seen along the contacts of some of the dikes transecting granitic rock. These minute quartz veinlets extending from the dike into the granitic rock penetrate between and surround the constituent crystals thus producing a ring-like pattern. It is possible that some of the water for these hydrothermal solutions came from the wall rocks.

TEXTURAL FEATURES OF A CONTAMINATED DIABASE DIKE

Most of the Tertiary diabase dikes which cut across Cornucopia Mountain are through-going ones without lateral terminations and are essentially free from inclusions. One dike, however, is unique in that it does end abruptly and that it is filled with inclusions adjacent to its termination. This dike, exposed high up (7250') on the steep walls of a glacial valley two miles west of the old townsite, has a width of 25 feet, a north-south trend and a 40° westerly dip. It transects quartz-dioritic country rock and is parallel to and locally in contact with one of the gold-quartz vein zones of the area. Fine grained chilled borders which are free from inclusions line both walls of the dike as well as its wedge-shaped northerly extremity. For a hundred feet south of here the central part of the dike contains so many angular xenoliths of granitic country rock, of silicified hornfels (which does not outcrop in the immediate vicinity) and xenocrysts of quartz and feldspar that it is lighter in color than the inclusion free main part of the dike to the south. Where the quartz xenocrysts are abundant the rock might easily be mistaken for a dacite were it not for the presence of reaction rims around the quartz grains.

Most of the xenoliths, which average four to five inches in size, are haphazard in their distribution in the central part of the dike, and show no preferred orientation. Some larger ($2 \times 1\frac{1}{2} \times \frac{1}{2}$ feet) granitic xenoliths are, however, roughly parallel to the walls of the dike. Thin sections of specimens taken along the borders of one of these slab-like granitic fragments show small (5 mm.) embayments of the diabasic matrix into the xenolith. Here small laths of plagioclase display a rough flow alignment. A few millimeters further into the xenolith the diabase forms a network of small veinlets which penetrate around the quartz and feldspar crystals of the granitic xenolith. About 10 mm. further into the xenolith these veinlets (0.1 to 0.5 mm. in width) show a gradual increase of secondary products with a concomitant decrease of pyrogenetic minerals, so that their composition is quite similar to that of the deuteric fraction locally abundant in the mesostasis of this dike. The secondary products are very fine grained in texture and consist chiefly of somewhat plumose aggregates of micropegmatitic intergrowths with numerous needle-like crystals of

tremolite, some grains of magnetite, minute flakes of a pinkish mica, a little secondary quartz, a few clear untwinned feldspar crystals and indeterminate grains of mafics and brownish micaceous minerals.

Although the initial fractures which became the pathways for the network of veinlets may well have been caused by the heat of the diabasic magma, the dominant mechanism appears to be one of replacement. The evidence for this is especially clear with respect to those of the deuteritic fraction. These veinlets show cusp-like indentations against the quartz grains with rather sharp contacts and no reaction rims, although along some parts of the boundaries the veinlets contain very minute disconnected lines of fine dust-like material. Along other parts of the boundaries, however, thin slivers of quartz protrude into the veinlets, and some grains of corroded quartz have bizarre shapes with two larger parts connected by a thin link. A few of the quartz grains are traversed by very minute (0.004 mm.) microveinlets of zeolite.

The plagioclase crystals of the xenolith also show similar replacement features where they are in contact with the veinlets, except that they are not as conspicuously embayed as the quartz grains and that most of them retain their rectangular shape but with rounded corners. Some of them show a penetration of the deuteritic fraction along the albite twinning planes with thin relics of the lamellae extending into the veinlets. Adjacent to the veinlets several of the feldspars display rough irregular patches of a mosaic-like or fretwork-like structure and some of the smaller ones are completely matted by a very fine (0.004 mm.) fretwork. A few of these are replaced by clear secondary feldspars in the form of clear, minute (0.02 mm.) rectangular or hollow square crystals.

Quartz grains and plagioclase crystals detached from the deuterically altered granitic xenolith are engulfed as xenocrysts by the igneous matrix. These xenocrysts display superimposed textural features that are clearly indicative of magmatic reactions at relatively higher temperatures than those of the deuteritic fractions. Similar features also occur with respect to xenocrysts of quartz and plagioclase distributed at random throughout the dike. The most noticeable of these features are the reaction rims of pyroxene around the quartz xenocrysts. The rims are from 0.1 to 0.2 mm. wide and the small, closely packed prisms of monoclinic pyroxene (probably diopside) are arranged radially, tangentially, or at other angles with respect to the quartz xenocrysts (Fig. 4). Most of these reaction rims include a few grains of magnetite and rarely minute euhedral crystals of a brown micaceous mineral. A few of the small pyroxene crystals penetrate slightly into the xenocrysts so that minute appendages of quartz extend into the reaction rims. In some thin sections there are roughly rounded aggregates of pyroxene crystals similar to

those in the reaction rims. Some of these may represent the complete magmatic replacement of a quartz xenocryst and others merely the outer part of a reaction rim.

Reaction rims of pyroxene do not occur around plagioclase xenocrysts, but other features produced by the magma are quite noticeable. These include rounded and irregular borders, and a finely fretted texture (0.01 mm.) which is coarser than that produced by the deuteritic fraction. The borders of some of these xenocrysts appear to have started to disintegrate into innumerable minute blocks (Fig. 4). Other plagioclase xenocrysts are partially or wholly covered by a fine fretwork; and where

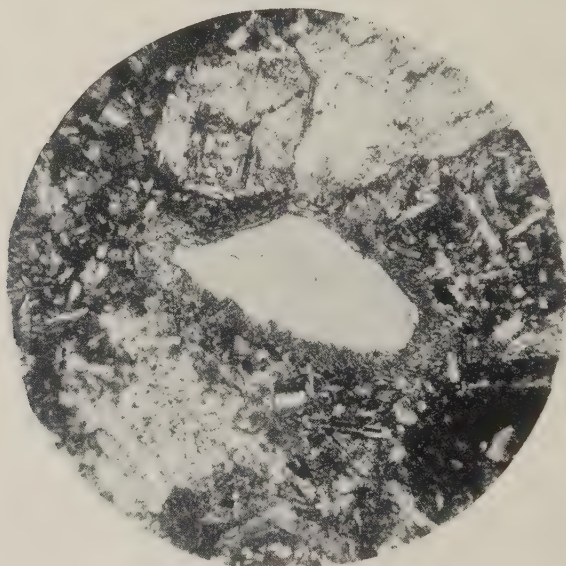


FIG. 4. Photomicrograph (plane light) showing xenocrysts of quartz and plagioclase from a thin section of a specimen from the contaminated dike. The 3 mm. quartz xenocryst (center) has a reaction rim of prisms of pyroxene. The xenocrystic plagioclase shows various stages of disintegration. Note the fretwork developed in one of the xenocrysts of plagioclase.

this occurs there is commonly a narrow (0.1 mm.) border of clear plagioclase similar in composition to the unaffected portions of the xenocrysts. Most of the plagioclases of the quartz-dioritic wall rock are in the oligoclase-andesine range of composition. The interstitial part of the fretwork in some of the plagioclase xenocrysts consists of extremely fine grains (0.002 to 0.06 mm.) of minerals of high refringence and birefringence with numerous small euhedral crystals of magnetite as well as some rod-like or staff-like forms of this mineral.

Some of the plagioclase xenocrysts show recrystallization effects such as the development of ragged porphyroblasts of monoclinic pyroxene, irregular interlocking patches which have a faint birefringence suggestive of feldspathic recrystallization, and the transformation of a few of the xenocrysts into a fine mosaic of clouded feldspar crystals. In contrast to this type of recrystallization, some small xenocrysts are changed into a checkerboard aggregate of several hundred subhedral, blocky, clear, twinned andesine (An₃₅) crystals averaging about 0.02 mm. square. Some of these are oriented at right angles to the extinction position of the original xenocryst, but most of them have an over-all conformity to the original crystal. Some of these checkerboard aggregates show a zigzag penetration of pyroxene in and around some of the small compact crystals. Where these small plagioclase crystals break away from the irregular borders of the aggregate, they merge into and become a part of the igneous matrix of the dike thus indicating assimilation without melting into a glass. In some of the plagioclase xenocrysts, however, the interstitial part of the fretwork contains several very small (0.1 to 0.01 mm.) irregular drop-like forms of a brown, isotropic, nearly opaque material which corrode the feldspar in minute cusps. They appear to be glass, and if so, represent a local incipient melting of the xenocrysts.

In addition to these effects of recrystallization, disintegration and assimilation of xenocrystic material, this dike displays many noteworthy features with respect to its conspicuous deuteric fraction. Deuteric products are not only noticeable in the mesotaxis, but also occur in irregular segregations, veinlets, and amygdules. The relationship of the mesotaxis to an amygdule is well shown in one thin section where a small (2 mm.) round aggregate contains numerous thin (0.7 mm. long) crystals of clear oligoclase and monoclinic pyroxene. Some of these pyroxenes are almost wholly replaced by granular magnetite. The material between these larger crystals is light brown in color and is peppered by innumerable minute mineral grains of high refringence and birefringence and magnetite in small crystals of dust-like or rod-like forms. Numerous microlites of oligoclase, some grains of pyroxene, minute plumose clusters of zeolites, patches showing faint ill-defined extinction are also present. One-half of the border of this round aggregate is clearly defined against the igneous matrix by several curved crystals of oligoclase (0.2 to 0.4 mm. long) whereas the other half, not as well marked, shows locally a merging of the deuteric fraction of the igneous matrix into the aggregate.

Another larger (5 mm.) round amygdule is well shown in a large thin section cut from an inclusion rich specimen taken near the hanging wall of the dike. The deuteric fraction filling this amygdule consists chiefly of a plexus of incipient feldspar microlites (0.3 mm. long) with some inter-

stitial zeolites and brown mesostasis which has a very faint birefringence and is filled with extremely minute opaque dust-like material. Within this amygdale is a smaller one (3 mm. in diameter) consisting of zeolites, and there are also several small (0.2 mm.) similar amygdules along part of the periphery of the host amygdale and some small irregular veinlets of zeolite. This section also contains a few small amygdules consisting chiefly of zeolites (Fig. 5).

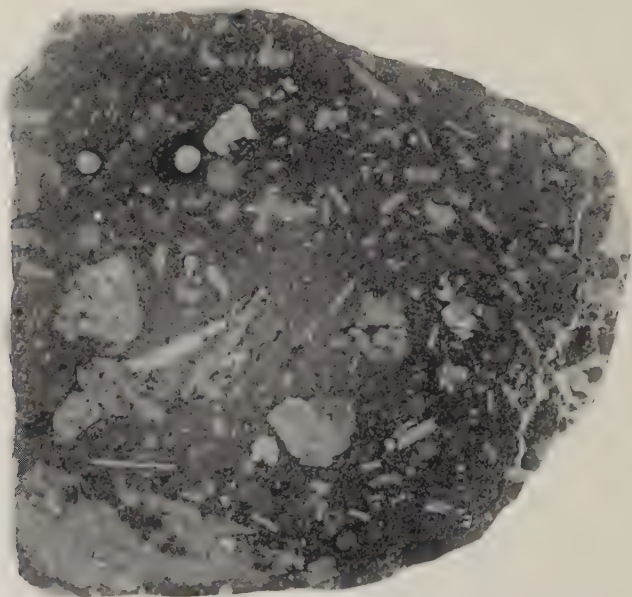


FIG. 5. Photograph (ordinary light) of a 6×7 cm. thin section of a specimen from the hanging wall of a contaminated diabasic dike. Note the irregular xenoliths of quartz granulite and siliceous hornfels which are recrystallized and partly disintegrated, and a few quartz xenocrysts surrounded by a narrow (0.1 mm.) reaction rim of pyroxene. Scattered through the section are numerous rounded or irregular patches of xenocrystic plagioclase in various stages of assimilation. In contrast are the sharp rectangular crystals of pyrogenetic labradorite. Note also the two round (light colored) amygdules of zeolites surrounded by an outer amygdale of the dark colored deuteritic fraction.

These features suggest that the deuteritic fraction was not only able to seep into a vesicle but was also locally capable of forming a vesicle which became filled with the final products of its crystallization sequence. At this late stage of crystallization the deuteritic material appears to have changed into a local hydrothermal solution. This is suggested by several other features such as the previously described vermicular veinlets, and the alteration of the pyrogenetic labradorite phenocrysts. Many of these

crystals show the penetration of zeolites along minute irregular fractures and cleavage cracks accompanied by varying degrees of replacement (Fig. 3).

SOME TEXTURAL FEATURES OF MAGMATIC ROCKS FROM VARIOUS LOCALITIES

In the Cornucopia diabase dikes the textural features of the orthomagmatic sequence of crystallization is, in general, quite distinct from that of the deuteric fraction. In many other basic igneous rocks which occur in much larger bodies, and, therefore, have a longer consolidation period, it is not always a simple matter to distinguish strictly magmatic textures from those of the crystallized deuteric fraction. Moreover, in some igneous masses, later extraneous metamorphism may impose quite different textural features and so make the textural pattern more complicated.

Sudbury, Ontario.

For example, many thin sections from specimens of the noritic part of the Sudbury, Ontario, intrusive show alteration to such an extent that only a few relics of the original mafics remain and even the plagioclases are partly seritized. The amount of micropegmatitic intergrowths as seen in some of these sections is far greater than that which might be expected from a normal crystallization sequence of a gabbroid magma. The normal sequence of crystallization is, however, illustrated by the textural features and mineral composition in a few specimens from certain localities near Sudbury. One of these specimens taken about 500 feet from the foot-wall of the intrusive body north of the Murry mine is a good example of the unaltered norite. It is a dark gray, medium grained crystalline rock consisting chiefly of clear tabular plagioclases (with a maximum length of 5 mm.) with interstitial mafics and a few flakes of brown biotite. In thin section it is seen to consist of about 58 per cent labradorite An₅₁, 23 per cent pyroxene and a little hornblende, 9 per cent alkalic feldspar, 6 per cent biotite, 3 per cent quartz, and 1 per cent opaques. The labradorite laths average 1 to 2 mm. in size, have a haphazard arrangement, and some of the larger ones show a few irregular fractures. Pigeonite (0.4 × 0.6 × 1 mm.) is the dominant pyroxene although some hypersthene is present. Some of the pyroxene is interstitial to the labradorite, some includes labradorite crystals showing a subaphetic texture, and some embays the plagioclase along cleavage cracks. Green hornblende forms rims around some of the hypersthene and also shows replacement textures against the pigeonite. Brown biotite mostly in solid patches exhibits boundaries indicating that it has partially replaced some of the horn-

blende, pyroxene, and plagioclase. A few flakes of biotite are embayed by quartz and a small one (0.2 mm.) showing some crystal faces is surrounded by quartz. The clear untwinned alkalic feldspar which has a small (—) 2v and is probably sanidine, looks like the quartz which, however, is optically positive. Both of these minerals occupy interstitial positions with respect to the other minerals. In contrast to most of the Sudbury "norite" this specimen contains only an extremely small amount of micropegmatitic and myrmekitic intergrowth: one very small crystal of plagioclase is partially replaced by a myrmekitic intergrowth and is surrounded by clear alkalic feldspar which is in an interstitial position to labradorite. The opaques are magnetite, pyrrhotite, and pyrite. One irregular aggregate of pyrrhotite (2 mm. in size) shows replacement cusps against the labradorite and also penetrates between some of the labradorite crystals. This aggregate of pyrrhotite is transected by a few minute (0.01 mm.) veinlets of micaceous minerals and is also partially rimmed by them. Another irregular aggregate (1.0×1.5 mm.) is entirely enclosed by biotite, but the borders between these minerals indicate that the biotite has been partially replaced by the sulphide. A few small grains of pyrite are rimmed with magnetite, and magnetite has partially replaced both biotite and hornblende developing the usual replacement textures.

The textural pattern of the earlier magmatic minerals of the Sudbury norite, namely, the labradorite and the pyroxenes, bears a close resemblance to that of some gabbros and many diabases especially with respect to the random orientation of the feldspars and the intergranular and in part subophitic texture (Fig. 6). Although some of the pyroxenes show small irregular penetrations into the labradorite, on the whole the boundaries between these minerals are clear cut and sharp. In contrast to this, some of the later formed minerals show distinct replacement textures against the earlier formed ones and the minerals last to form, namely the alkali feldspar; the quartz and the sulphides occupy interstitial positions comparable to that of the deuteric fraction as seen in some diabases. This feature suggests that these minerals crystallized at a time when the rock was a close-knitted net of earlier formed magmatic minerals and that they represent the final stage of consolidation.

Yogo Peak, Little Belt Mountains, Montana

The textural features of the later consolidation products of a basic magma are well shown in some of the shonkinitic rocks of central Montana. At the classical locality of Shonkin Sag, shonkinite occurs both above and below a relatively narrow band of microsyenite and a narrower one of syenite pegmatite. The upper shonkinite consists chiefly of euhedral augite (up to 5 mm. in size) in a random arrangement and a

few crystals of euhedral to subhedral olivine some of which are enclosed by the pyroxene, numerous block-like forms of biotite with some showing crystal faces and some surrounding the earlier formed minerals. The mesostasis consists for the most part of a partly altered confused aggregate of micropegmatitic intergrowth, alkalic feldspars, some nephelite and a little fibrous stilbite. Embayment of biotite by the mesostasis and cusp-like indentations into the biotite are indicative of replacement by the mesostasis. The mineral association as well as the textural pattern of the mesostasis suggest that it formed at a very late stage in the consoli-

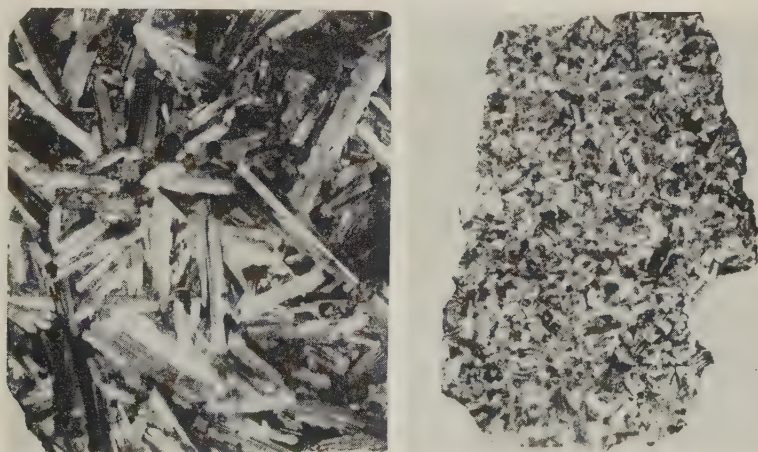


FIG. 6. Photographs (crossed polarized light) of a 4×4.5 cm. thin section of Duluth gabbro and a 3×4 thin section of the Sudbury, Ontario, norite. Note the criss-cross pattern of the labradorite crystals.

dation of the rock. It is of interest to note that in this shonkinite the mafics, particularly the augites, display marked euhedralism, whereas the interstitial feldspathic material is mostly quite anhedral.

Somewhat similar textural features are present in the shonkonitic border phases of the Yogo Peak chonolithitic intrusive mass in the Little Belt Mountains of north central Montana. As described by Weed and Pirs-son (1898), this mass is some five miles long and about one mile wide, the western end making the peak. Here the rock is a shonkinite which, as one proceeds eastward, changes into rocks of dioritic, monzonitic, syenitic character, to what has been termed a porphyritic granite (Fig. 7). A similar gradation from shonkinite to granite is noticeable from the eastern border westward. In the summer of 1949 the writer collected specimens one-tenth of a mile apart from the West peak of Yogo Mountain to the central part of the intrusive mass.

At the western border the rock is nearly black, coarse-grained, and

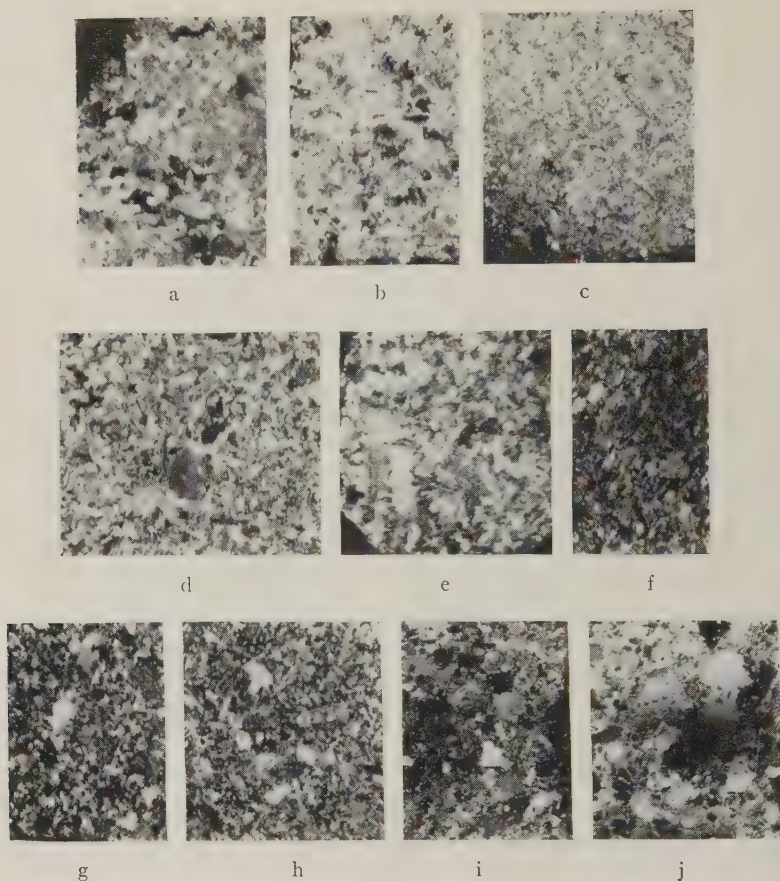


FIG. 7. Photographs (crossed polarized light) of ordinary thin sections from Yogo Peak to illustrate the gradual change from shonkinite to granite porphyry, as follows: a) Yogo West Peak; b) 0.3 Mile East; c) 0.4 Mile East; d) 0.5 Mile East; e) 0.6 Mile East; f) 0.7 Mile East; g) 0.9 Mile East; h) 1.1 Mile East; i) 1.3 Mile East; j) 1.4 Mile East.

consists chiefly of numerous plates of biotite (5 to 10 mm.) showing a poikilitic texture, pyroxene, a few grains of which are euhedral, some greenish rounded grains of olivine and some interstitial feldspar. As seen in thin section, euhedral to subhedral pyroxenes (2 to 5 mm. in size) with an occasional remnant of olivine are included poikilitically by subhedral to anhedral blocky brown biotite. All these minerals are included by large (10 to 15 mm.) clear anhedral plates of feldspar mostly orthoclase with a little andesine, and oriented at random. Magnetite in small irregular grains or rod-like forms is for the most part scattered through the mafics but some is concentrated in altered relics of what was probably

originally olivine. In one part of a large thin section from this rock there is an irregular cluster (10×5 mm.) of prismatic hornblende crystals arranged haphazardly with interstitial carbonate. A few of the pyroxenes are embayed by biotite, some of them show the penetration of biotite along minute fractures, and others contain minute flakes of biotite. Some of the altered mafics are rimmed with green biotite and a few of the feldspars are slightly clouded by sericitic material. Although this shonkinite from the western peak of Yogo Mountain is characterized by euhedral pyroxenes like the one from Shonkin Sag, it differs in that instead of a confused aggregate of interstitial feldspathic material, large plates of feldspar, mostly orthoclase, enclose the earlier-formed minerals.

The igneous rock three-tenths of a mile east of West Peak is slightly finer-grained than that at the western border and similar although slightly different textural features are revealed in the thin sections. The mafics are not as large, most of the pyroxenes averaging 1 mm., with a few olivine crystals of the same size. The large plates of orthoclase also include several small crystals of oligoclase, some of which show ragged borders and embayments indicative of replacement by the alkalic feldspar.

At four-tenths of a mile to the east the igneous rock is finer-grained and not quite as dark in color as that to the west. Several altered mafics (5 mm. in size), probably originally olivine, are noticeable and on exposed surfaces these have weathered out giving a pseudovesicular appearance to the rock. Interstitial feldspathic material is more conspicuous than it is in the previous specimens. Thin sections show that the mafics, with the exception of the olivine, are smaller, ranging in size from 0.1 to 1.0 mm. Most of the mafics have ragged borders and embayments indicative of replacement by the feldspathic mesostasis which is an aggregate of small (0.1 to 0.3 mm.) subhedral oligoclase enclosed in an interlocking aggregate of anhedral alkalic feldspar. A few of the mafics, especially the olivines, are rimmed by small flakes of green mica and contain numerous grains of released magnetite.

At a half mile east of the western border, the igneous rocks become finer-grained and gray in color. Interspaced with the mafics are numerous clear subhedral crystals of feldspar. Biotite is the most noticeable of the mafics, but some hornblende and pyroxene are recognizable in the hand specimen. As seen in a large thin section under a wide field microscope, clear subhedral plagioclase with rectangular outlines or groups of plagioclase crystals are in part surrounded by smaller, earlier-formed mafic crystals which appear to have been pushed aside by the growing feldspar. It is clear that the mafics formed earlier than the plagioclases because they are embayed by the feldspars which also contain a few inclusions of

mafics. The mafics are hornblende, pyroxene, (diopsidic augite) and biotite, and the plagioclase is oligoclase An₂₇. A late-forming mineral, occurring interstitially, is anhedral alkalic feldspar which appears to have partly replaced all the other minerals including the oligoclase. Instead of being clear like the plagioclase, this feldspar is turbid, being filled with dust-like inclusions of minute mafic crystals and a few euhedral crystals of apatite. There is also a little interstitial quartz with replacement borders against the other minerals. Some interstitial magnetite which embays the earlier minerals is in turn embayed by the alkalic feldspar.

The igneous rock for the next two-tenths of a mile is similar both megascopically and microscopically to that just described except that the plagioclases, instead of having a random orientation, show some degree of parallelism. This preferred orientation of the plagioclases is quite evident on large cut surfaces of the rock or in large-sized thin sections. It is probably the record of late stage magmatic flowage. This rock also contains a few small (5–15 mm.) irregular inclusions consisting chiefly of an aggregate of biotite with interstitial plagioclase liberally sprinkled with grains of magnetite. In the central part of one of these inclusions is a small segregation of pyrite.

At nine-tenths to one and one-tenth of a mile east of the western border of the Yogo intrusive mass, the igneous rock does not show any flow structure, becomes lighter in color due to the diminution of mafics, and has about the same grain size. Thin sections reveal that orthoclase is the dominant mineral with subordinate oligoclase, some crystals of which are partly or almost wholly replaced by orthoclase. Some of the larger (up to 3 mm.) crystals of orthoclase are composite ones consisting of several interlocking anhedral. These usually contain inclusions of oligoclase or phantom relics of this mineral. Some microperthite is also present. The mafics constitute about ten per cent of the rock and consist of hornblende, pyroxene, and biotite. A few relic pyroxenes are heavily mantled with hornblende. Some of the mafics are embayed by the feldspar, but a few of the hornblendes show some crystal faces against them. Most of the mafics contain grains of magnetite, and a few of them include flakes of brown biotite and occasional grains of sphene. This rock also contains from three to five per cent of interstitial quartz which also penetrates and partly replaces the earlier minerals. Although the overall texture of this rock could be described as hypidiomorphic granular, the irregular interlocking boundaries of the feldspars as well as the prevalence of replacement textures strongly suggests a crystalloblastic growth in the deuteric fraction of an originally basic magma. Relics of pyrogenetic minerals such as the pyroxenes confirm the magmatic lineage of the rock as shown by the field evidence.

Less than one-half mile farther to the east the igneous rock has changed into what has been called a porphyritic granite. At a distance of 1.3 miles from the western border of the intrusive mass, it has a light cream color and numerous conspicuous crystals of orthoclase (5 to 7 mm.) in a finer-grained matrix of feldspar, grains of quartz, and a few mafics consisting of biotite and hornblende. In thin section the texture of this rock is seen to be very different from those in the western part of the Yogo intrusive mass. Large (3 to 7 mm.) crystals of alkalic (orthoclase) feldspar, some of which are slightly microperthitic, compose about 50 per cent of the rock. A few of these larger feldspars are compound crystals where several anhedral with different orientations and irregular boundaries form one crystal. There are also a few conspicuous anhedral of quartz ranging in size from 1.3 to 2.2 mm. Interstitial to the larger alkalic feldspars are numerous anhedral (0.5 mm.) of alkalic feldspar, subhedral oligoclase 1–2 mm., perthitic anhedral (1.5 mm.), ragged relict hornblende (2 mm.) partly replaced by biotite, blocky biotite (1 mm.), anhedral quartz (1 mm.) a few relics of augite and an occasional crystal of sphene, zoisite, and magnetite. The mesostasis of this matrix consists of a mosaic-like aggregate of anhedral quartz (0.2 mm.) in rounded or polygonal forms, and anhedral feldspar (0.2 mm.) orthoclase and oligoclase. The mesostasis embays a few of the larger quartz anhedral which are also traversed by minute (0.01 mm.) veinlets of alkali feldspar. The larger alkalic feldspars include some of the small subhedral oligoclase and partly replaced oligoclase crystals. These larger crystals in turn appear to have been invaded by the small quartz grains of the mesostasis which appear along their borders as advanced islands of replacement. Many of these rather complex textural features are suggestive of crystalloblastic growth. The field relations as well as the relics of pyrogenetic minerals, however, clearly indicate that this granitic rock is of a magmatic lineage.

The Yogo Mountain intrusive mass is transected by numerous aplitic dikes. One of these cutting the shonkinitic rock near the western border is 2 cm. wide and contains a few small xenoliths which, aside from being elongated not quite parallel to the walls of the dike, do not appear to have been changed by the aplitic magma. In thin section this small dike is seen to consist chiefly of anhedral to subhedral alkalic feldspar (1 mm.) some of which is perthitic, a few small (0.7 mm.) crystals of oligoclase many of which are partly replaced by alkalic feldspar. Minor constituents are interstitial quartz which has replaced some of the feldspars; relict hornblende usually with a little magnetite, a few flakes of biotite some of which show some crystal faces, and a few interstitial grains of sphene. This rock is similar texturally to some of the dioritic facies, but does not have nearly as many mafics.

Osgood Mountains of Northeastern Nevada

Some thirty miles to the east of Winnemucca, Nevada, in the Osgood Mountains, is a small granodioritic intrusive body described by Hobbs and Clabaugh. This intrusion has an outcrop form of a figure eight with a north-south dimension of six miles, a width of two miles in the bulges and only 1000 feet in the narrow connection between the wider parts. The country rock consisting of interbedded argillites and limestone dips away from the igneous body, with dips of 40° to 60° on the eastern side and nearly vertical on the western side. The contact aureole is relatively narrow and irregular, and most of it is characterized by the usual lime silicate minerals. Locally small parts of the contact aureole have been granitized.

The intrusive rock is light-colored, medium to coarse grained with a few large (1 cm.) plagioclase crystals showing rectangular outlines in a matrix of smaller feldspars, mafics, and interstitial quartz anhedral. Hornblende and biotite are recognizable, and some of the biotite shows nearly euhedral basal sections. Under the microscope the larger plagioclases are seen to be oligoclase (An₂₇) and to include numerous smaller (2 cm.) plagioclase crystals. There are a few ragged remnants of pyroxene and hornblende. Hornblende also occurs as small euhedral crystals (0.1+ to 0.7 mm.). Brown biotite is in a blocky rather than a shredded form, and there are a few larger (2.5 mm.) euhedral basal sections of biotite, some of which include small (0.5 to 1.0 mm.) plagioclase crystals. Platelike forms of anhedral alkalic feldspar (sanidine) surround the smaller crystals of plagioclase, biotite, and hornblende. Interstitial quartz anhedral range from 1 to 2 mm. in size. There is no question as to the intrusive nature of this igneous rock, and its magmatic lineage is clearly shown by the relict mafics. Mobility of the final stage of consolidation is indicated by the euhedralism of the late-forming biotite.

SOME TEXTURAL FEATURES OF METASOMATIC ROCKS

The textural features of magmatic rocks are, in general, the result of a crystallization sequence from an initially fluid magma to a nearly solid medium. The magma is usually limited by the boundaries of the intrusion and with falling temperature solidification and crystallization proceed from the borders inward. In the late magmatic or deuteric stage of crystallization the residual magmatic solutions may replace some of the earlier formed pyrogenetic minerals thus producing textural features somewhat similar to those found in metasomatic rocks.

Crystallization by metamorphism takes place in an essentially solid medium with first rising, then falling, temperatures or even under condi-

tions of undulating temperature. It also takes place progressively from one part to another of a pre-existing rock mass rather than throughout the whole mass at one time. This mass may consist of any kind of rock and therefore will present a variety of physical and chemical properties which, in part, are accountable for the diversified results of metamorphism. Another factor, which is sometimes most significant, is the effect of stress not only previous to, but during metamorphism.

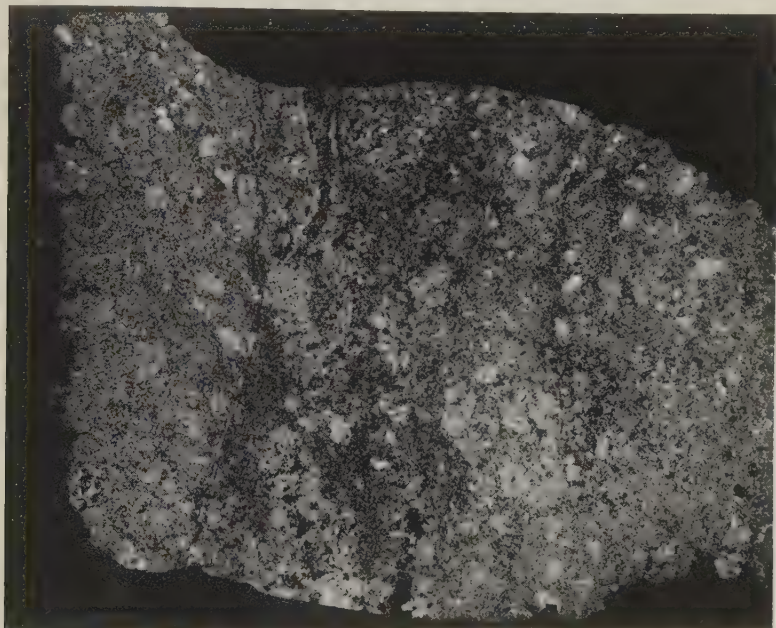


FIG. 8. Photograph (crossed polarized light) of a 5×7 cm. thin section from a specimen of hornblende granulite from the lowest adit, Cornucopia, Oregon. Note plagioclase porphyroblasts in an early stage of development.

Metasomatic rocks are those metamorphic rocks which have been formed by the addition of material carried by emanations or solutions through minute fractures or permeable zones. This material aided by a rising temperature reacts with the constituents of the rocks and new minerals are formed. Unlike the crystallization of a magma, no orderly sequence obtains and individual minerals commonly show various stages of crystalloblastic development. Within the area of a single section, porphyroblasts of the same mineral species may be displayed in several stages of development from initial amoeboid forms through those with ragged crystal outlines with pronounced sieve structure to nearly euhedral crystals which are practically free from inclusions (Fig. 8).

The character of the inclusions seen in these crystals depends upon the nature of the original material, or the stage of metamorphism preceding the development of the porphyroblasts. In contrast to inclusions in minerals in igneous rock, they are not governed by a definite crystallization sequence as is illustrated by included round grains of quartz in calcic plagioclase or hornblende. Relics of a former schistosity; i.e., helicitic structure or bands of platy or acicular minerals in porphyroblasts are cogent proof of metamorphic origin. Some plagioclase porphyroblasts occurring in a hornblendic granulitic rock in the vicinity of Cornucopia, Oregon, show rings of inclusions which have apparently been pushed outward by the growing crystal. In some of these porphyroblasts zonal structure coincides with the disconnected rings of inclusions. This suggests that the absorption of mafic inclusions has had some influence on the chemical composition of the plagioclase, and may be one of the causes of the oscillatory zonal structure. In other feldspathized rocks from this area it is not uncommon to see plagioclase porphyroblasts partially rimmed by the finer-grained schistose groundmass, thus indicating the effect of crystal pressure so commonly seen with respect to garnet porphyroblasts.

Where porphyroblasts are grouped together, the terms *glomeroblastic* for crystals of the same mineral or *cumuloblastic* for groups of different minerals were suggested by the writer some twenty years ago (Goodspeed, 1937). Some glomeroblastic aggregates seem to represent an intermediate stage in the formation of larger porphyroblasts. In the earlier stages of crystalloblastic growth, these groups are merely aggregates of crystals; then with further development the individual crystals coalesce to form a single larger crystal. The outline of this crystal is apt to be quite irregular with sharp re-entrant angles resulting from the growing together of several individual crystals. With continued development, this irregular outline may be modified or may disappear with the result that it becomes a single crystal. In this crystal, however, the twinning and other structures of the smaller individual crystals may persist and thus reveal its composite nature. In the later stages of development these features may be nearly obliterated so that only a hazy mottling gives a clue as to its former composite character. Although glomeroblastic aggregates are superficially similar to glomerophyric aggregates in igneous rocks, their metasomatic origin is evident where they occur in rocks of unquestionable metamorphic origin (Fig. 9).

For example, near Buffalo Hump, Idaho (Goodspeed, 1942), glomeroblastic aggregates of plagioclase occur in a fine-grained biotite schist. Most of the smaller aggregates (1 cm.) have irregular, ragged gradational boundaries with the schist, and some show helicitic structure. Others are

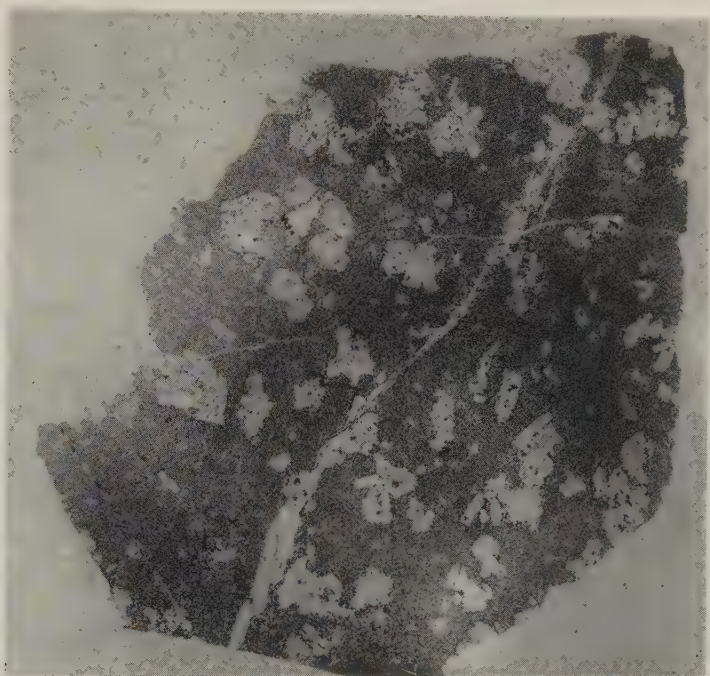


FIG. 9. Photograph (ordinary light) of a 5×6 cm. thin section of hornblende granulite, Cornucopia, Oregon. Note the glomeroblastic plagioclase with inclusions.

grouped in a radial fashion, have rounded outlines like miniature incipient orbicules, and display transitions from glomeroblastic aggregates to small but fully formed orbicules in the schist. Some of these appear to have developed around a core of schist.

These orbicules occur in zones about 50 feet wide and parallel to the schistosity. They range in size from less than an inch to several inches in diameter. Most of them are fairly regular ovoids with their major axes up to three inches in length. Although most of the orbicules are closely clustered, some are scattered in the adjacent fine-grained quartz biotite schist (Fig. 10). Around these scattered orbicules is a secondary schistosity which suggests that the growing orbicule pushed aside the minerals of the schist in a manner similar to a garnet porphyroblast. Where the orbicules are closely clustered, they are in the form of rounded polygons apparently due to mutual interference at the time of crystalloblastic growth. Some orbicules actually merge with one another. Where the orbicules are close together, the matrix loses its schistosity, becomes slightly coarser grained, and shows a marked increase in crystalloblastic feldspar. Even here, however, thin tabular relics of schist grade into the

coarser grained matrix which otherwise has the appearance of a fine-grained granitic rock. Thin sections, especially large sections, of this orbicular rock reveal some of the complex textural features of glomeroblastic growth. In general the orbicules are plagioclase porphyroblasts or sheaf-like aggregates of andesine (An₃₄) arranged in cross section something like a four-leaf clover with the intervening spaces occupied by crystals or aggregates of different orientation. The larger plagioclases commonly include other small plagioclases in various positions and ir-

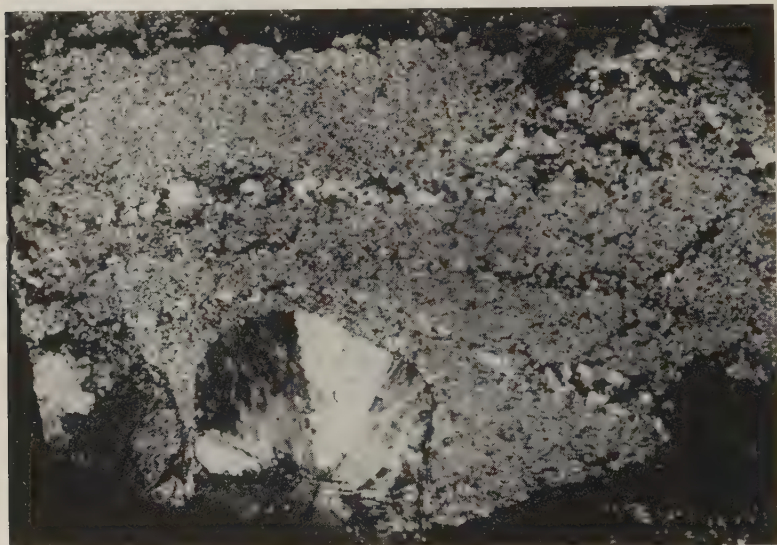


FIG. 10. Photograph (crossed polarized light) of a 6×9 cm. thin section of a biotite schist from Buffalo Hump, Idaho. Note the glomeroblastic plagioclase in the form of a 25 mm. orbicule.

regular patches of unstriated feldspar. In some orbicules later formed oligoclase-andesine (An₂₈₋₃₀) transects and embays the earlier andesine (An₃₆₋₃₇) thus partly replacing it. Some of these earlier plagioclases have distinct outlines and may show rough zonal structures where the central part is more calcic than the outer zone. In other orbicules the plagioclase has a mottled appearance suggesting that it originally consisted of many individuals. Polished specimens of some of the orbicules display minute vugs which show crystal faces of the constituent minerals. It is clear from both the field and petrographic evidence that these orbicules are the result of crystalloblastic growth rather than magmatic crystallization.

Hornblende in rocks of magmatic derivation is commonly seen to have

surrounded and replaced earlier pyrogenetic pyroxene. In rocks of metasomatic origin hornblendes may be seen in various stages of development: First, several small separate hornblende crystals have an over-all similar orientation as is shown by their extinction position; then a larger, very ragged inclusion-filled hornblende apparently is the result of a partial coalescence of the small crystals; finally, a more compact homogeneous crystal is formed. Although these hornblendes display a few crystal forms, some ragged borders and inclusions which persist give a clue as to their mode of origin.

Large biotite porphyroblasts may also develop from glomeroblastic aggregates of small flakes of biotite which appear to have coalesced into a larger individual with fairly definite crystallographic outlines. Such a large biotite porphyroblast will commonly have a poikiloblastic texture and crenulated border and show thin extensions into the surrounding groundmass.

Where porphyroblasts in any stage of development are fairly evenly distributed throughout a rock, it is an indication of either isochemical recrystallization or widespread permeability with respect to additive emanations or solutions. As the porphyroblasts develop, and as their number increases, the original material of the rock is either absorbed or pushed aside by them so that under favorable conditions they finally form a dominant part of the rock. Although the textural features of feldspathized rocks bear a superficial resemblance to those of magmatic rocks, they differ in that the later-formed feldspars are usually more euhedral than the earlier ones and the relict material consists of metamorphic minerals which commonly show some vestige of earlier metamorphic structures.

Later structural features such as fractures, fracture and shear zones, and brecciated zones are commonly a controlling factor with respect to subsequent feldspathization. This relationship is quite prominent in the Cornucopia hornfels and is well shown in many places along the low level 6400-foot adit. Here zones of complex fracturing display varying degrees of feldspathization from single fractures lined with porphyroblasts to areas of feldspathization bounded by fracture and shear planes.

Where fracture planes are closely spaced the feldspathized body has a tabular dike-like form. These have been termed replacement dikes, and in the Cornucopia area they can be seen in various stages of development; namely, (1) the initial stage where closely spaced parallel fracture surfaces form a zone along which incipient mineralization has taken place; (2) partial replacement of the intervening screens of wall rock by appropriate minerals; and (3) finally, a complete recrystallization replacement with the obliteration of nearly all traces of the wall rock ma-

terial. In some replacement dikes these traces of wall rock material are merely thin trains of relict minerals which retain their original orientation; in others, septa of the country rock extend from wall to wall or are in the form of projections into the dike (Fig. 11).

Replacement dikes have a wide range in mineral composition depending upon the character of the wall rock and the penetrative solutions. Where permeable zones are a controlling factor, they are likely to have gradational borders whereas control by two distinct parallel fractures or

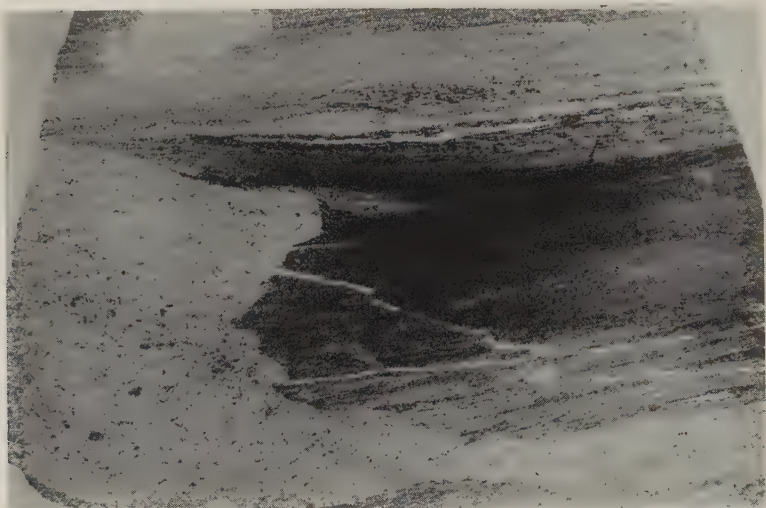


FIG. 11. Photograph (ordinary light) of a 6×9 cm. thin section showing a contact of a replacement dike with a hornblende biotite schist. Note the thin lines of mafics extending from the schist into the dike, which also contains streaks of mafics having, in general, the same orientation as the mafics in the schist.

shears commonly produces sharp borders. In his study of aplite and pegmatite dikes in Nigeria, King (1948) gave an excellent description of non-dilation dikes with sharp borders. In contrast to igneous dikes, replacement dikes do not show field evidence of dilation such as appropriate offsets of wall rock units. Chilled borders are lacking, and there is no progressive change in grain size from the borders inward. There may, however, be local changes in grain size within the dike resulting in an uneven crystalloblastic texture with some minerals or aggregates of mineral much larger than others. These larger crystals lack the regularity in size that is usually seen in the phenocrysts of a porphyritic igneous rock.

Some replacement dikes contain inclusions which look like xenoliths, but the lack of magmatic reaction products, the presence of schistosity

similar in attitude to that of the wall rock as well as the textural and structural features of the dike do not favor a magmatic interpretation. As these inclusions become feldspathized with the development of porphyroblasts similar to the feldspars of the dike, they assume a shadowy appearance showing but little contrast to their crystalloblastic matrix and for this reason have been termed skialiths (Goodspeed, 1948).

Where initial fracturing of the country rock has been in roughly parallel planes, elongated relict fragments are the rule; whereas in cataclastic zones unoriented blocky relics are common. In the latter case, relict pieces of country rock may be seen in various stages of transformation from well defined fragments to hazy skialiths hardly discernible from the crystalloblastic matrix. It is also not uncommon to see two relatively large parts of a fragment connected by a thin link, thus negating the possibility of magmatic flowage of the matrix. Such fractures coupled with the crystalloblastic textures of the matrix warrants the term replacement breccias for these occurrences.

Replacement breccias are common along many of the granitic hornfels contacts where there are also much larger inclusions of hornfels in the granitic rocks. These large included blocks could, of course, be interpreted as remnants of roof pendants or as the result of magmatic stoping. Many field and petrographic data however do not favor this interpretation. Detailed plane table maps of several critical areas in the vicinity of Cornucopia, on a scale of 200 feet to the inch, show several large inclusions some of which are in two parts connected by a relatively thin link of hornfels. Another significant feature of some of these inclusions is the preservation of structural features such as remnants of plunging folds. These structures might be interpreted as the result of forceful magmatic intrusion, but such an explanation is very doubtful since the strike of these folds is nearly at right angles to the main granitic mass and some of them are truncated by the granitic rocks, and others appear to be undisturbed relics surrounded by the granitic rock.

Most of these larger blocks of hornfels have gradational borders with the granitic rock, and some have been changed to a migmatite consisting of numerous smaller irregular relict fragments interspersed with granitic rock. These fragments usually range from two to ten cm. in size and display the schistose texture which has an over-all general alignment despite the separation of the fragment. This alignment is, no doubt, inherited from the attitude of the original schistosity in the hornfels. In thin sections, the fragments exhibit typical crystalloblastic textures with hornblende and biotite in a parallel arrangement. There are also some porphyroblasts of plagioclase and hornblende and narrow bands of anhedral aggregates of quartz. In the granitic parts of the thin sections

there are trains of mafics, hornblende, and shreds of biotite and elongated narrow aggregates of quartz anhedral (0.02 to 0.3 mm.) which retain the same general direction that they have in the hornfels fragments. Biotite has been partly replaced by feldspars as is shown by selective embayments along cleavages and minute linear inclusions of biotite in the feldspars. Many of the hornblendes can be seen to have been derived from the fragments, and as they extend further into granitic matrix, some are bent around the plagioclase crystals in the matrix. Here the feldspars range in size from 0.5 to 3.0 mm. and consist mostly of subhedral to anhedral oligoclase-andesine (An₃₁) with a few anhedral crystals of orthoclase. Many of the plagioclases show various kinds of zoning and are quite complex, as though made up of several individual crystals grown together. Minute grains of metamorphic minerals, zoisite, garnet, and epidote are included in some of the plagioclases and also occur interstitially with grains of magnetite and sphene.

Other exposures along the numerous hornfels-granitic contacts at Cornucopia show similar textural features, and a few show a complete gradation from schistose hornfels into a gneissic rock which, in turn, grades into a granitic rock with a directionless texture. Locally, the granitic rocks display variations in texture and composition ranging from dioritic types to leucocratic and aplitic varieties. In general, however, as seen in hundreds of thin sections of specimens from all parts of the area, the granitic rock is a quartz-diorite with a seriate crystalloblastic fabric of anhedral to subhedral oligoclase-andesine (2+4 mm.) with interstitial anhedral quartz, feldspar, and ragged flakes and shreds of brown or green biotite. This mineral as seen in most thin sections is positioned around the larger plagioclase crystals in such a way as to suggest that it had been pushed aside by the growing feldspar (Fig. 12).

Many of the plagioclases are complex crystals and some show oscillating zoning. In G.S.A. Memoir 52, R. C. Emmons (Chapter 9, p. 111) states: "Oscillatory zoning in plagioclase of an igneous rock is interpreted in Chapter 4 to reflect the role of liquid in crystallization. However, thin sections, especially large ones, may show the percentage of oscillatorily zoned plagioclase to be low enough to suggest that liquid has played a locally prominent but generally minor role. This irregular distribution of significant zoning has been observed in metamorphic rocks and in those which show brecciation. It is interpreted as a dilatant effect characteristic of metasomatic granites." Not only is there an irregular distribution of oscillatory zoned plagioclase in the Cornucopia granites but this feature is also noticeable in some of the porphyroblasts of feldspathized schistose hornfelses.

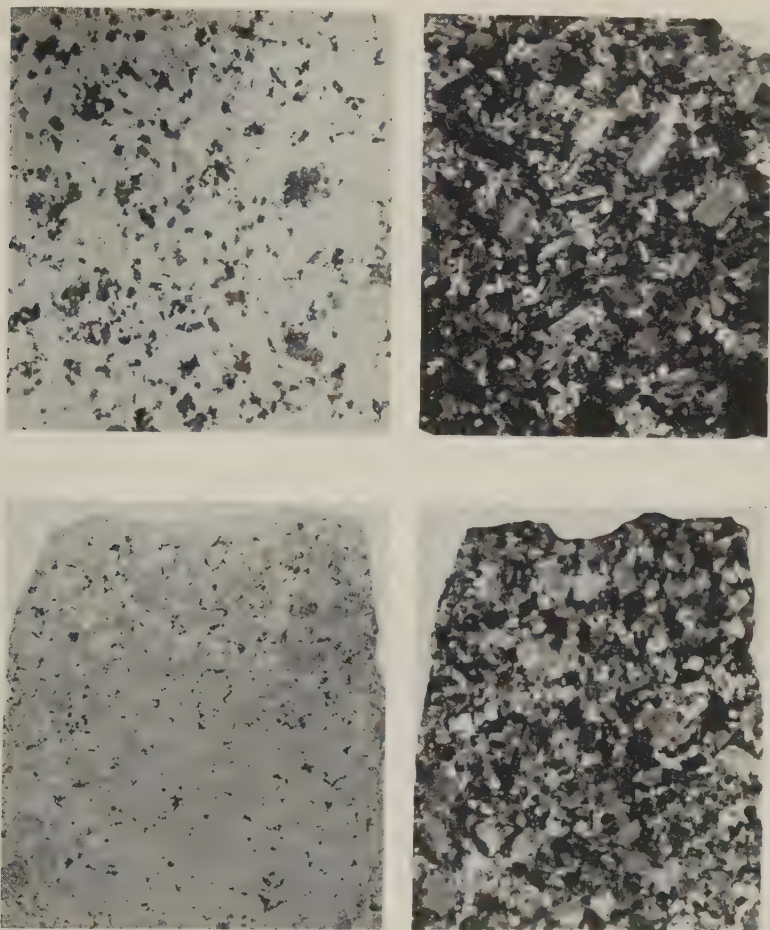


FIG. 12. Photographs (in ordinary light and crossed polarized light) of granitic rock; (1—above) 4×4 cm. thin section of a specimen from Boulder, Montana; (2—below) 3×4 cm. thin section from the lowest adit at Cornucopia, Oregon. Note that in (1) the early formed plagioclase is subhedral to euhedral, the quartz fills spaces between the other minerals, and the biotite is blocky. In (2) the later plagioclase is subhedral, most of the quartz is in rounded grains and the shredded biotite surrounds the plagioclase.

The preceding descriptions have had reference, in general, to metasomatized rocks where there has been a considerable change in grain size and mineral composition from the original material to the final product. If, however, the pre-existing rock happened to be an arkose, which already has a mineral composition and grains size comparable to a granite, the degree of transformation is much less. All stages of the transforma-

tion of an arkose to a granite have been well described by Coombs (1950) in a paper entitled "Granitization in the Swauk Arkose near Wenatchee, Washington."

Most of the examples of metasomatic rocks just described have been taken from relatively small rock bodies. Some large-size granitic masses occur in the central parts of metamorphic terranes. They usually grade into adjacent gneisses and schists which in turn blend into less metamorphosed rocks. Such features have been described by P. Misch (1949) in three papers entitled "Metasomatic Granitization of Batholithic Dimensions," in which he demonstrates the relationship of high-grade regional metamorphism to synkinematic granitization and its relation to static granitization. Since many kinds of metamorphic rocks are involved, wide variations in textural features are to be expected although the dominant pattern is a crystalloblastic one. In synkinematically granitized rock the minerals, especially the mafics, usually display a preferred orientation.

RELICT DIKES AND RELICT PSEUDODIKES

In some parts of the Cornucopia area are small dike-like bodies which can be seen to be definitely pre-granitic in age, because they are cut by granitic veinlets or penetrated by crystalloblastic extensions of the granitic rock. Their irregular or shadow-like borders and locally gradational contacts are indicative of partial replacement by the surrounding granitic rock. They also have attitudes nearly the same as the replacement dikes in the nearby hornfels, and this feature, coupled with their petrographic similarity to some of the replacement dikes, suggests that they were originally replacement dikes in the hornfels before it was transformed into granitic rock and that already having been changed from hornfels into a more granitic-like rock, they were able to resist further change and remain as relict dikes (Goodspeed, 1955). Similar relics of long thin tabular masses of schistose hornfels surrounded by the granitic rock also occur in this area. Since these are merely dike-like remnants of metamorphic rock, they are termed relict pseudodikes. The presence of relict dikes or relict pseudodikes suggest that the granitic rock was emplaced by a gradual transformation of the country rock because it is difficult to see how such thin tabular bodies could have escaped being disturbed by magmatic intrusion.

MOBILIZATION--RHEOMORPHISM--NEOMAGMAS

The textural features of rocks from some occurrences in the vicinity of Cornucopia are somewhat paradoxical in that they display crystalloblastic features common to metamorphic and metasomatic rock and

yet have some features, including fluxion textures, which are characteristic of magmatic rocks (Goodspeed 1952). For example, a small leucocratic dike well exposed in the bed of Pine Creek, two miles west of Cornucopia, reveals features indicative of both replacement and mass flowage. A coarse gneissic rock is transected by this dike, and at one place near one wall delicate remnants of mafic minerals extend from the gneiss into the dike and retain their original alignment which is nearly at right angles to the dike. Along the opposite border, however, these mafic remnants appear to have swirled away from the wall and tend to become parallel to it. This occurrence is called a mobilized replacement dike. In other somewhat similar dikes, metasomatized material has apparently flowed as a neomagma into a dilated fissure. These are called rheomorphic dikes.

The textural features of some breccias also, like those of rheomorphic dikes, present the apparently contradictory evidence of a crystalloblastic pattern and fluxion textures. Some breccias in the Cornucopia area show, in part, the characteristics of a replacement breccia and also locally a flow alignment of mafic minerals in the matrix. Others also show a flow alignment of the fragments. Likewise, breccias in other localities; i.e., in the northern Cascade Mountains of Washington or at Sudbury, Ontario, the textural evidence of flowage is not only confined to the matrix but is emphasized by the rounded or drawn-out elongated character of the fragments. These have been termed rheomorphic breccias so as to differentiate them from static replacement breccias or from those of orthomagmatic origin, and to emphasize the interpretation that they represent the mobilization of a partly metasomatized rock mass (Goodspeed 1953).

CONCLUSIONS

Some rocks display field relations and textural features which present a clear picture as to whether they have resulted from the crystallization of a magma or from the transformation of pre-existing material. For other rocks where the textural features emphasize only the final stages of crystallization, detailed field evidence is of primary importance for petrogenetic interpretations.

The textural features of the rocks from some basic intrusive bodies delineate the changes in crystallization from a fluid magma to an essentially solid medium. Although different conditions modify the textural patterns, the sequence of crystallization is mainly an orderly one, as, for example: (1) High temperature minerals such as olivine, pyroxene, and calcic plagioclase are early in the sequence, and if plagioclase happens to form earlier than the pyroxene a characteristic criss-cross pattern is the

usual result. (2) Although initially these minerals may be quite euhedral they will, if a slow rate of cooling prevails, be transformed in accordance with Bowen's reaction principle so that they become merely relics. (3) In the later stages of crystallization, minerals lower down in the reaction series may replace, surround, or form interstitially to, the earlier minerals. Some of these later minerals may be euhedral and others anhedral, dependent upon immediate environmental conditions. (4) The very late magmatic or deuteric stage is characterized by low temperature minerals which commonly show crystalloblastic and replacement textures. (5) Finally, hydrous minerals like the zeolites may appear. Some intrusions such as the Yogo Peak body in the Little Belt Mountains of Montana show a range of crystallization of an orthomagma from a shonkinite to granite porphyry.

The textural features of metasomatic rocks are similar to those of metamorphic rock, in that crystalloblastic textures are dominant and show a crystallization sequence which is not an orderly one like that of a magmatic rock. Minerals, such as feldspars, which may in part be due to the addition of material, can be seen in a single thin section, in forms indicative of their development from initial amoeboid ones to nearly euhedral crystals. The recrystallization of a pre-existing rock is indicated by helicitic structures, the presence of rounded grains of quartz or relics of typical metamorphic minerals like kyanite. Another indication that crystallization took place in an essentially solid medium is where earlier minerals seem to have been pushed aside by later ones. For example, in some granitic rocks ragged shred-like flakes of biotite surround large plagioclase crystals somewhat similar to the way they enclose garnet porphyroblasts in a metamorphic rock. In the field, metasomatic rock masses commonly show widespread gradations into the surrounding metamorphic rocks, in contrast to the limited extent of metamorphic aureoles around intrusive bodies.

Intrusive field relations, however, do not necessarily mean that the rock body has been formed by the crystallization of an orthomagma or that it is of orthomagmatic lineage. This holds for some granitic bodies which are clearly intrusive and which are commonly termed igneous. Their textural features, however, may indicate one of three possibilities with regard to their mode of origin, namely: (1) Direct crystallization or differentiation from an orthomagma; (2) Mass flowage of a late magmatic or deuteric fraction; that is, a deuteromagmatic origin; (3) Mobilization or mass flowage of metasomatized material, namely, a neomagmatic origin. This last possibility is perhaps the most difficult to recognize because it depends in part upon the presence of relics of metamorphic minerals.

With respect to some relict minerals, the question arises as to whether they are of metamorphic or of magmatic origin. For example some intrusive granitic bodies contain relics of pyroxene, and since this mineral is common to magmatic rocks the interpretation of a magmatic line of descent is usually proposed. Pyroxene, however, is also found in some metamorphic rocks. Therefore the mere presence of pyroxene does not furnish a definitive answer to this question. What is needed are data to show, if possible, whether a particular mineral was originally crystallized from a magma or was formed by rising temperature in a solid rock. Perhaps laboratory experiments in the future will disclose whether there are any slight but significant differences in artificial rock-forming silicates produced from melts as compared with similar ones formed by heating material that is essentially solid.

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A STRUCTURAL STUDY OF IDDINGSITE FROM NEW SOUTH WALES, AUSTRALIA

G. BROWN AND I. STEPHEN, *Rothamsted Experimental Station,
Harpenden, Herts., England.*

ABSTRACT

Iddingsite from New South Wales, Australia, is polycrystalline and consists of goethite and a layer lattice silicate. In the alteration of olivine to iddingsite the original lattice of close-packed oxygens appears not to have been greatly disturbed and the changes have occurred by the movement of cations within small regions to form microcrystals of the alteration products. The parallel alignment of the components explains why it behaves optically as a single crystal.

INTRODUCTION

Iddingsite is the name generally given to the deep reddish brown to ruby red lamellar minerals commonly found in basaltic and allied rocks as an alteration product of olivine (Winchell, 1951). The origin, occurrence, composition, and physical properties of iddingsite were first described in detail by Ross and Shannon (1926), and subsequent studies include those by Tomkief (1934), Edwards (1938), and Bogue and Hodge (1940). Since this investigation was begun Ming-Shan Sun (1957) published the results of x -ray powder examinations of iddingsites, and concluded that goethite was the only substance present in the crystalline state, and that other substances whose presence was shown by chemical analysis were largely amorphous. Wilshire (1958), using a similar technique, found that iddingsites were largely smectite-chlorites with goethite always present; quartz and calcite are common and talc and mica rare constituents.

The present study was undertaken to investigate iddingsite more adequately, and in particular to find the reason for its optical homogeneity, and also to obtain information on a naturally occurring alteration process. The material used was obtained from an olivine-basalt of Tertiary age from near Lismore in northern New South Wales, Australia (Norrish, 1952), and grains were isolated that were favourable for study by optical and single crystal x -ray diffraction techniques.

PETROGRAPHY

The rock is a porphyritic olivine-basalt with phenocrysts of labradorite and olivine in a medium-fine grained groundmass of abundant laths of a less basic plagioclase, augite, ilmenite, nontronite and apatite. The phenocrysts of labradorite range up to 1 cm. in size, while those of the olivine attain a length of about 2 mm. Some of the olivine grains are completely transformed to iddingsite, but usually unaltered cores of oli-

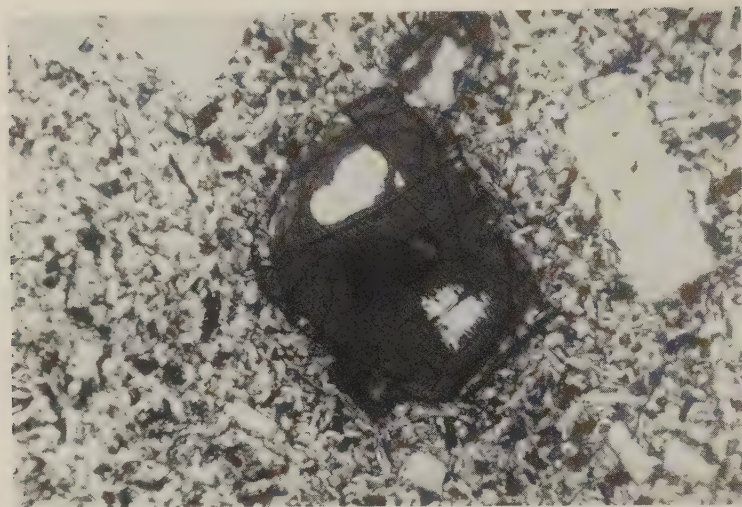


FIG. 1. Olivine-basalt, Lismore, N.S.W., Australia, showing an iddingsite pseudomorph with cores of remnant olivine. Magnification: $\times 25$.

vine are preserved within the iddingsite with a fairly sharp boundary, which, in places, may be somewhat ragged, suggesting reaction (Fig. 1). The pseudomorphs are normally rimmed by a thin zone of yellowish brown or greenish cryptocrystalline material. It is probable that the iddingsite is of deuteric origin, as the associated minerals show little signs of weathering, and, in particular, the labradorite phenocrysts are very clear and unaltered.

OPTICAL PROPERTIES

In thin section the iddingsite shows a lamellar habit, with one well developed cleavage, and two subsidiary cleavages at right angles to each other and to the principal cleavage direction: extinction is parallel to the cleavage traces and to the extinction directions of the remnant olivine. Pleochroism is distinct with Y and Z deep orange brown to reddish brown and X light orange brown to yellowish brown.* The majority of the individual grains isolated after crushing the rock are tabular in habit due to the presence of the well developed cleavage. These cleavage fragments are deep reddish brown and are essentially non-pleochroic. The refractive indices vary slightly but most of the grains have $\alpha = 1.67$ – 1.68 and $\gamma = 1.71$ – 1.72 , with the birefringence approximately 0.04. The optic

* A few of the olivine grains show an alteration product having olive green tints in its pleochroic scheme in addition to shades of brown: morphologically the material is similar to iddingsite, but this variant has not been studied.

TABLE 1. ζ VALUES (CuK α) AND DESCRIPTION OF LAYER LINES ON PHOTOGRAPH OF IDDINGSITE ROTATED ABOUT Y OPTIC VIBRATION DIRECTION

ζ	Axial repeat in Å	Nature of reflections	
		Broad	Sharp
0.0	—	Many, strong	present
.075	20.6	Few, weak	
.15	10.3	Many, strong	present*
.23	6.7	Few, weak	
.29	5.3	Few, strong	
.31	5.0	Many, strong	present*
.38	4.1	Few, weak	
.46	3.35	Few, medium	present*
.52	2.98	Few, weak	
.58	2.68	Few, weak	
.62	2.50	Few, weak	present*
.76	2.00	Few, weak	present*

* These sharp spots do not lie precisely on the same layer lines as the broad spots, but are close to them.

sign is negative with moderate optic axial angle ($2V$ 30–40°): the dispersion is distinct, $r < v$.

The grains selected for x -ray analysis were optically homogeneous, showing the emergence of the acute bisectrix normal to the plates, and giving a centred biaxial interference figure from which the directions of the optical constants could readily be determined.

X-RAY ANALYSIS

A preliminary single crystal rotation photograph of a grain (0.25 \times 0.25 \times 0.05 mm.) rotated about the Y optic vibration direction showed well developed layer lines, but the pattern was complex. Table 1 summarizes the ζ (CuK α) values, the corresponding axial repeat distance, and the nature of the layer lines. The majority of the reflections were broad, but on or near a number of layer lines were very small sharply defined spots: comparison with an olivine b -axis rotation photograph showed that the sharp spots could be attributed to olivine.

The layer lines of broad spots with $\zeta = n(.075)$ were provisionally attributed to one mineral, and those with $\zeta = n(.29)$ from the axial repeat distance and hexagonal arrangement of the inner reflections were attributed to a layer lattice mineral rotated about the $a = 5.35$ Å axis. A powder photograph of a few crushed grains suggested that goethite was present, and also showed a reflection $d = 15.6$ Å which could be the basal reflection of the layer lattice silicate.

The powder pattern was poor, and to obtain further information, Weissenberg and oscillation photographs were taken with the Y optic direction as the rotation axis. As many of the reflections were weak and broad the Weissenberg photographs showed only the stronger spots, which were insufficient for complete characterization. From a complete set of 5° oscillation photographs reciprocal lattice nets were obtained for each of the twelve levels of the broad spot pattern, and also for the zero, first and second levels of the material giving the sharp spots: these were constructed by plotting the co-ordinates of the reflections, which were 5° arcs rather than points. The reciprocal nets showed that the symmetry of the entire diffraction pattern is *mmm*. Complete sets of oscillation photographs were also taken about the two other orthorhombic axes, corresponding to the X and Z vibration directions, and the reciprocal nets were constructed for all broad spot levels.

All the applied criteria indicate that the olivine is being transformed into a material comprising goethite in three orientations and a layer lattice silicate with layer stacking disorder.

Olivine

The cell dimensions $a=4.8$, $b=10.3$, $c=6.0$ Å and space group *Pbnm* of the material giving the sharp spots confirmed its identification as olivine. Bragg and Brown (1926) gave $a=4.755$, $b=10.21$, $c=5.985$ Å and space group V_h^{16} ($\equiv Pbnm$) for olivine. Powder photographs of crushed iddingsite grains with included olivine gave for the latter $d(130)=2.779$ Å corresponding to the composition Fo (80%) determined by the method of Yoder and Sahama (1957). The olivine axes are oriented with $a\parallel X$, $b\parallel Y$ and $c\parallel Z$ of the iddingsite.

Goethite

The goethite was identified from the cell dimensions $a=4.6$, $b=10.0$, $c=3.0$ Å, determined from the layer line separations and the reciprocal nets, and the systematic absences, $h0l$ absent for $(h+l)$ odd, $0kl$ absent for k odd, which are consistent with the space group *Pbnm*. Goldsztaub (1931) gave $a=4.64$, $b=10.0$, $c=3.03$ Å and the space group *Pbnm* for goethite. The absences were derived from rather limited data, reflections with $2\theta > 90^\circ$ being too diffuse to index with certainty. The observed intensities, however, gave additional evidence, for 110 the strongest reflection is the strongest goethite reflection and all the strong observed intensities correspond to strong goethite reflections.

Most of the goethite is aligned parallel to the original olivine with

$$a_G\parallel a_F; \quad b_G\parallel b_F; \quad c_G\parallel c_F^\dagger$$

† The subscripts G, F and L refer to goethite, olivine and the layer lattice silicate respectively.

The goethite in this orientation accounts for the reflections occurring on the $\zeta=n$ (.15) layer lines (Table 1). When account was taken of the reflections due to the layer lattice silicate it was found that those remaining could be attributed to goethite oriented so that

$$a_G \| a_F; \quad b_G \| [0\bar{1}3] \text{ and } [0\bar{1}3]_F; \quad c_G \| [0\bar{1}1] \text{ and } [0\bar{1}1]_F$$

In these orientations only a few medium or weak reflections could be observed, and these correspond to the strong goethite reflections. Fig. 2 illustrates the three orientations of the goethite cells in relation to the original olivine. From the intensities of the 110 reflection from goethite in the three orientations about 80 per cent appeared to be in parallel orientation with the olivine and only 10 per cent in each of the other two orientations. From the width of the reflections the size of the individual goethite crystals is estimated to be 200–300 Å.

Layer lattice silicate

This component accounts for the layer lines $\zeta=n$ (.29) (Table 1). The reciprocal lattice consists of a single row of reciprocal lattice points passing through the origin, and reciprocal lattice rods parallel to the point row arranged in a hexagonal pattern around the central row. The repeat period along the point row corresponds to a spacing of 15.6 Å and the cell side of the hexagonal array is 5.35 Å. The stacking of the layers is highly disordered, the only observable regularity in addition to the parallel and equidistant stacking of the silicate layers being the parallel alignment of the a -axes of the $(\text{Si}_2\text{O}_5)^{2-}$ layers which make up the half layers of the mineral. Table 2 gives ξ values, derived values of h^2+hk+k^2 and hk indices for the rods which are continuous except $11l^*$ and $22l^*$ which have zero intensity at $l^*=0$. The values of h^2+hk+k^2 confirm the hexagonal nature of the array.

The a dimension corresponds closely to the value found for trioctahedral layer lattice silicates, e.g. chlorite $a=5.33\text{--}5.37$ Å (Steinfink, 1958) and vermiculite $a=5.33$ Å (Mathieson and Walker, 1954). The repeat distance along the point row corresponds to the layer thickness of smectites in the air-dry condition when saturated with divalent cations. Unlike most smectites, however, the basal spacing could not be increased by treatment with water, ethylene glycol or glycerol. The layer thickness is reduced to 10 Å by heating at 500° C. for one hour. These properties appear sufficient to identify the material as a mineral of the vermiculite or smectite group, but obviously a more detailed examination is desirable. The layer lattice mineral of iddingsite has been identified by Wilshire as a smectite-chlorite. However, the collapse to a layer thickness of 10 Å on heating, both of the majority of the specimens studied by Wilshire and of those examined in the present study, does not fully support his con-



FIG. 2. Projection along a^* of reciprocal lattice of goethite showing its three orientations. Heavy continuous line shows goethite cell in same orientation as olivine, axes b_1^* and c_1^* .

Dotted line and broken line show goethite cells in subsidiary orientations with axes $b_2^*c_2^*$ and $b_3^*c_3^*$ respectively

Δ , \bullet and \circ denote observed reciprocal lattice points for orientations 1, 2 and 3.

\circ are reciprocal lattice points where corners of the differently oriented cells coincide, and thin continuous lines show large reciprocal cell of the close packed oxygens.

TABLE 2. ξ VALUES (CuK α) AND hk INDICES (HEXAGONAL) OF RECIPROCAL LATTICE RODS OF LAYER LATTICE MINERAL; IDDINGSITE ROTATED ABOUT X OPTIC VIBRATION DIRECTION

ξ	ξ^2	$h^2 + hk + k^2$	hk
.33	.11	1	10
.58	.33	3	11
.65	.42	4	20
.88	.77	7	21
1.00	1.00	9	30
1.14	1.30	12	22
1.19	1.42	13	31
1.44	2.07	19	32
1.53	2.34	21	41

clusions. Although slight atomic re-arrangement takes place, the basal spacing of chlorite layers remains near 14 Å until recrystallization takes place (Brindley and Ali, 1950), and interstratified smectite-chlorite would be expected to give intensity maxima between 10 and 14 Å after heating.

The orientational relationships between olivine and the layer lattice silicate are as follows:

$$a_L \parallel [010], [0\bar{1}3] \text{ and } [0\bar{1}3]_F; \quad c_L \parallel [100]_F$$

The use of a and c here is not intended to imply a true unit cell, but only to denote a structural unit.

DISCUSSION

It has been shown that the transformation of olivine into iddingsite involves the production of new crystalline phases, and it is pertinent here to consider more closely the relationships of these new materials to the parent olivine.

Olivine and goethite belong to the same space group $Pbnm$ and the cell dimensions have the following relations:

$$a_F \simeq a_G; \quad b_F \simeq b_G; \quad c_F \simeq 2c_G$$

Both structures are based on sheets of oxygen in hexagonal close-packing parallel to (100), which, considering oxygens only, leads to the equivalence of the directions $[010]$, $[0\bar{1}3]$ and $[0\bar{1}3]_F$ at angles of 120° ; similarly $[001]$, $[0\bar{1}1]$, and $[0\bar{1}1]_F$ are equivalent. Structurally only slight adjustments of the oxygen framework are required in the change from olivine to goethite; the cation replacements and rearrangements could easily take place by ionic diffusion. The three observed orientations of goethite are explained by the equivalence of the three directions mentioned above. The preponderance of goethite in parallel orientation to olivine suggests, however, that the three directions are not exactly equivalent and that the cation arrangement in olivine exerts considerable control over that of the derived goethite. It is interesting to note that the triple points, where the corners of the differently oriented goethite cells coincide, outline a hexagonal cell $a = 2.9$ Å, $c = 4.6$ Å, which corresponds to the unit cell of hexagonal close-packed oxygens (Fig. 2).

The alteration of olivine to a layer lattice silicate involves some breaking up of the close-packed oxygen framework of the olivine. Nevertheless in a study of what is in effect the reverse reaction, the alteration on heating of layer lattice silicates (chlorite and serpentine) to olivine, Brindley and Ali (1950) and Brindley and Zussman (1957) have shown that the majority of the Si—O bonds remain unbroken, only slight rotations of

tetrahedra and migrations of Si being required to effect the alterations. In these studies the following dimensional and orientational relations were established between the layer lattice mineral and the resultant olivine.

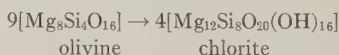
$$2a_L \simeq b_F \text{ and } 2b_L \simeq 3c_F \ddagger$$

$$[100]_L \parallel [010] \text{ and } [013]_F; \quad [010]_L \parallel [001] \text{ and } [011]_F$$

The same mechanism is proposed for both mineral groups, the transformation being visualised as occurring layer by layer, one layer of chlorite giving two layers of olivine and one layer of serpentine yielding one layer of olivine. The same dimensional and orientational relations are found in the present study if the transformation from hexagonal to face-centred orthorhombic axes for the layer lattice silicate is made, and it appears that alteration takes place by a mechanism similar to that postulated above.

It is difficult to make any statements about the chemical changes involved in the absence of knowledge of the proportions of goethite and the layer lattice silicate in the iddingsite. Although the iddingsite occupies the same volume as did the olivine, its density (ca. 2.8) is considerably less than that of olivine (3.4–3.5) determined for Fo (80%) from the graph given by Winchell (1951, p. 500). Material has obviously been lost in the change, but as the system is not a closed one the overall loss is the resultant of both gains and losses. If the change is considered to occur layer by layer, one olivine layer is equivalent to one goethite layer, but the alteration to the layer lattice silicate is more complicated.

From the relationship $4(ab)_L \simeq 3(bc)_F$ and $c_L \simeq 3c_F$, an alteration without volume change involves nine cells of olivine giving four cells of the layer lattice silicate. If the silicate is considered as being a chlorite (this having the densest packing in the inter-layer region and therefore requiring most material), and ignoring isomorphous substitution, the equivalence in volume is found to be such that



From these approximations it is evident that large amounts of Mg are expelled from the regions which alter to the layer lattice silicate, and from the regions which alter to goethite all the Mg and Si are lost while Fe and H are added. These conclusions, loss of Mg and addition of Fe and H are similar to those arrived at by Ross and Shannon on the basis

‡ The $a = 5.3 \text{ \AA}$ and $b = 9.2 \text{ \AA}$ axes of layer lattice minerals lie in the plane of the layers; $b = a\sqrt{3}$ due to the pseudo-hexagonal or hexagonal nature of the layers.

of chemical analysis. The suggestion by Ming-Shan Sun that the ratio of Fe_2O_3 to $\text{H}_2\text{O}+$ represents the composition of goethite in iddingsite seems unlikely. Obviously the layer lattice silicate will contain considerable amounts of $\text{H}_2\text{O}+$ and it is improbable that it would accept Mg^{2+} and entirely reject Fe^{3+} .

The most unusual feature of iddingsite is that, although polycrystalline and consisting of two components, grains exhibit the optical properties of a single crystal. Describing the rocks of the Eureka district, Nevada, Iddings (1892) discussed the alteration product of olivine (later named iddingsite) and stated that "the resultant mineral from its optical properties is evidently not a confused aggregate, but a crystallographic individual, with parallel orientation of all its parts, for the extinction of light is the same throughout and the interference figure that of a doubly refracting crystal." As has been shown by the present investigation, this homogeneity in optical properties occurs because the small crystals of both components are strictly oriented throughout a single grain. The parallel alignment arises from the nature of the alteration, the products of which inherit, goethite completely and the layer lattice silicate partly, the oxygen framework of the original olivine. The same alignments of the major components probably holds for all iddingsites which show coherent optical properties.

ACKNOWLEDGMENTS

The authors are indebted to Miss Sylvia Reeves for drawing Fig. 2, and to Mr. V. Stansfield for the photomicrograph.

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DELRIOITE, A NEW CALCIUM STRONTIUM VANADATE FROM COLORADO*

MARY E. THOMPSON AND ALEXANDER M. SHERWOOD,
U. S. Geological Survey, Washington 25, D. C.

ABSTRACT

Delrioite, $\text{CaO} \cdot \text{SrO} \cdot \text{V}_2\text{O}_5 \cdot 3\text{H}_2\text{O}$, was found on the dump at the portal of the Jo Dandy mine, Montrose County, Colorado. The mine is in a vanadium-uranium deposit in the Salt Wash member of the Morrison formation of Late Jurassic age. The mineral occurs as radial aggregates of pale yellow-green fibrous acicular crystals. Unit cell data for delrioite are not available, but the x-ray powder diffraction pattern is unique. The d -spacings of the strongest lines are, in Angstroms: 6.5 (S), 4.95 (MS), 4.40 (M), 3.55 (M), 3.45 (MS).

Delrioite is biaxial negative, with indices of refraction $\alpha=1.783$, $\beta=1.834$, $\gamma=1.866$, all ± 0.003 , 2V medium to large. It is pleochroic with X colorless, Y pale yellow, and Z slightly deeper yellow. The hardness is about 2; the measured specific gravity is 3.1 ± 0.1 .

The chemical analysis shows, in per cent: CaO 13.30, SrO 24.50, V_2O_5 46.00, $\text{H}_2\text{O}(+)$ 5.64, $\text{H}_2\text{O}(-)$ 9.24, SiO_2 1.30, total 99.98.

Delrioite is named for A. M. del Rio, who first found vanadium (erythronium) in North America.

INTRODUCTION AND ACKNOWLEDGMENTS

Delrioite was found on the dump at the Hummer portal of the Jo Dandy mine in Paradox Valley, Montrose County, Colorado, in 1955, by Edwin Sharp, a summer field assistant, of the U. S. Geological Survey. The specimen was given to E. B. Gross of the U. S. Atomic Energy Commission, Grand Junction, Colorado for identification, who submitted it to M. E. Thompson for further study. We wish to thank Mr. Sharp and Mr. Gross for supplying us with the specimens of this new mineral.

Delrioite is named for A. M. del Rio (1764-1849), who in 1801 announced the discovery of a new element, erythronium. The announcement was later withdrawn, as he had become convinced that erythronium was really chromium. Erythronium was later shown to be identical to vanadium, the new element described by Sefstrom in 1830. It is, therefore, considered appropriate to name this new vanadium mineral for one of the earliest American mineralogists, whose share in the discovery of vanadium is usually forgotten.

This work is part of a program being conducted by the U. S. Geological Survey on behalf of the Division of Raw Materials of the U. S. Atomic Energy Commission.

OCCURRENCE

Delrioite occurs as an efflorescence on sandstone on the dump at the Hummer portal of the Jo Dandy mine in Paradox Valley, Montrose

* Publication authorized by the Director, U. S. Geological Survey.

County, Colorado. The Jo Dandy mine is located in a vanadium-uranium deposit in the Salt Wash sandstone member of the Morrison formation of Late Jurassic age. The occurrence as an efflorescence at the dump suggests that the new mineral is one of the soluble oxidation products of the lower valent vanadium minerals found in the Jo Dandy mine. The new mineral has so far been found only at the one locality, probably because an adequate supply of strontium usually is not available in the ground waters at other localities.

Another vanadium mineral found on the same dump is metarossite, $\text{CaV}_2\text{O}_6 \cdot 2\text{H}_2\text{O}$. The two minerals look very much alike, and might easily be confused, unless laboratory tests are made.

X-RAY, PHYSICAL, AND OPTICAL PROPERTIES

Delrioite occurs as pale yellow-green aggregates of fibrous acicular crystals. The aggregates approximate single crystals closely enough so that optical measurements can be made, but the fragments were found, in a preliminary study by x-ray rotation and Weissenberg methods, evidently to be twinned, and in addition, to consist of tiny fibers that are somewhat rotated with respect to each other about their mutual axis of elongation. This type of disorder has been observed in other fibrous vanadates (Barnes and Qurashi, 1952). The x-ray rotation photographs established, however, that the repeat distance along the fiber axis is 2 times 3.65 \AA , a distance typical of fibrous vanadates.

Delrioite is light yellow green, but exposed surfaces of the mineral may be darker green, probably because of photoreduction of a small part of the vanadium (V). The hardness is about 2; the luster is vitreous to pearly. The specific gravity, obtained by measuring several small fragments together on a Berman density balance, is 3.1 ± 0.1 .

The new mineral is biaxial negative; $2V$ is medium to large; $2V$ calculated from the indices of refraction is $78\frac{1}{2}^\circ$. The fibers exhibit parallel extinction, and the γ index is parallel to the elongation. Indices of refraction and pleochroism are as follows:

$\alpha = 1.783 \pm 0.003$	X colorless
$\beta = 1.834 \pm 0.003$	Y pale yellow
$\gamma = 1.866 \pm 0.003$	Z slightly deeper yellow

Delrioite has a unique x-ray powder pattern, for which d -spacings are given in Table 1. The mineral was photographed with Ni-filtered Cu radiation ($\lambda = 1.5418 \text{ \AA}$), in a Debye-Scherrer type camera of 114.59 mm. diameter.

CHEMISTRY

A sample of delrioite weighing about 100 mg. was purified by hand-picking for the spectrographic and chemical analyses. The spectrographic

TABLE 1. X-RAY POWDER DIFFRACTION DATA FOR DELRIOITE

CuK α (Ni filter), $\lambda=1.5418 \text{ \AA}$, camera diameter 114.59 mm. Measurements not corrected for film shrinkage. Cut-off 15 \AA^*

I estimated	$d \text{ (\AA)}$	I estimated	$d \text{ (\AA)}$
W	7.2	W	2.96
S	6.5	MW	2.80
F	6.2	M	2.69
MS	4.95	W	2.57
W	4.74	W	2.51
M	4.40	MW	2.18
M	3.55	F	2.12
MS	3.45	F	1.93
F	3.35	W	1.89
W	3.27	F	1.85
F	3.18	M	1.80
M	3.04	W, broad	1.77

S strong, M medium, W weak, F faint

* Two faint lines were recorded with the diffractometer at 8.30 \AA and 7.71 \AA .

analysis, by K. V. Hazel of the U. S. Geological Survey, showed over 10 per cent Si and V, 1 to 10 per cent Sr and Ca, and <1 per cent Mg and Fe. The large amount of Si was due to admixed quartz, which was mostly removed from the sample before the chemical analysis.

The chemical analysis, by Alexander M. Sherwood, yields the empirical formula $\text{CaO} \cdot \text{SrO} \cdot \text{V}_2\text{O}_5 \cdot 3\text{H}_2\text{O}$ (Table 2). This formula indicates that the mineral is a pyrovanadate, but the suggestion has been made (H. T. Evans, Jr., written communication, 1958) that the conditions of formation would more likely produce a metavanadate, in which case the

TABLE 2. CHEMICAL ANALYSIS AND CALCULATED COMPOSITION OF DELRIOITE
Alexander M. Sherwood, analyst

	Analysis (per cent)	Analysis recalculated to 100 per cent	Atomic ratios	$\text{CaO} \cdot \text{SrO} \cdot \text{V}_2\text{O}_5 \cdot 3\text{H}_2\text{O}$
CaO	13.30	13.5	0.241	14.14
SrO	24.50	24.8	0.239	26.26
V_2O_5	46.00	46.6	0.256	45.96
$\text{H}_2\text{O}(+)$	5.64	5.7	0.316	4.55
$\text{H}_2\text{O}(-)$	9.24	9.4	0.522	9.09
SiO_2	1.30			
Total	99.98	100.0		100.00

formula might be written as $\text{CaSrV}_2\text{O}_6(\text{OH})_2 \cdot 2\text{H}_2\text{O}$. The x-ray powder diffraction pattern, too, bears some resemblance to a synthetic calcium metavanadate (Daphne Ross and Richard Marvin, U. S. Geological Survey, oral communication, 1957).

The mineral is readily soluble in water, giving a yellow-green solution. Upon evaporation, however, a different (unknown) phase is obtained.

The empirical formula is parallel to that of pintadoite, $2\text{CaO} \cdot \text{V}_2\text{O}_5 \cdot 9\text{H}_2\text{O}$, a preliminary description of which was published by Hess and Schaller (1914). Unfortunately, available specimens of pintadoite seem to be nothing more than faintly stained sandstone so that it has not been possible to compare the two minerals directly.

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NARSARSUKITE FROM SAGE CREEK, SWEETGRASS HILLS, MONTANA*

D. B. STEWART, *U. S. Geological Survey, Washington 25, D. C.*

ABSTRACT

Narsarsukite, $\text{Na}_2\text{TiSi}_4\text{O}_{11}$, is described from the second of three localities in the Sweetgrass Hills, Montana. Narsarsukite is associated with manganpectolite, quartz, calcite, and galena in veins cutting an orthoclase-albite-aegirine-quartz hybrid rock in a zone adjacent to a syenite stock. Optical properties of the minerals and an indexed x-ray powder-diffraction pattern of narsarsukite are given.

This occurrence of the assemblage aegirine-quartz-narsarsukite is similar to other occurrences of narsarsukite in the Sweetgrass Hills and at Narsarsuk, Greenland. It is concluded that the formation of narsarsukite depends upon the reaction of emanations from alkalic magmas with siliceous wallrock under conditions such that a high partial pressure of oxygen is present.

INTRODUCTION

The Sweetgrass Hills in north-central Montana near the border of the United States and Canada consist of three related groups of stocks. The geology of the region is summarized by Ross (1950). East Butte, the group of stocks that contains the narsarsukite localities, is the largest, most easterly, and most complex of the three groups of stocks. The alkali-rich igneous rocks from East Butte are described in detail by Kemp and Billingsley (1921). The more than a dozen closely-spaced syenite and trachyte plutons of East Butte have domed and metamorphosed the sedimentary rocks that border them. Siliceous metasedimentary rocks adjacent to the plutons contain small deposits of iron and copper, and veins in marble close to the stocks contain fluorite and quartz. Brecciated highly metamorphosed rocks that have been intricately intruded by syenite make up a hybrid rock that was mapped separately by Ross (1950). The hybrid rock borders only the largest stock and is related to it. All the narsarsukite occurrences are in the hybrid rock, and quartz-bearing veins outside the hybrid rock do not contain narsarsukite.

The first description of narsarsukite from the Sweetgrass Hills was by Graham in 1935. Graham's locality is in the upper drainage basin of Halfbreed Creek, in the center of sec. 24, T. 36 N., R. 4 E. Another occurrence of narsarsukite was discovered by the writer in 1950; additional specimens were collected in September 1951. The new occurrence, about one mile northeast of Graham's locality, is in the NE $\frac{1}{4}$ of sec. 19, T. 36 N., R. 5 E., on the bottom and west bank of Sage Creek at an elevation of 5300 feet. The exposure is only about 6 feet by 6 feet, and is 150 feet upstream from the place where the wagon trail to the divide between Mount Royal and Mount Brown crosses Sage Creek. The Madi-

* Publication authorized by the Director, U. S. Geological Survey.

son limestone crops out about 200 feet above the locality on the east side of Sage Creek canyon. Specimens from this locality are deposited at the U. S. National Museum under Museum number 106424.

Narsarsukite has also been noted in veinlets in blocks of hybrid rock in the float on the southwest side of Mount Royal (W. T. Pecora, oral communication, February 1958).

PETROGRAPHY

Hybrid Rock

Greenish-gray aphanitic hybrid rock is the host rock for the narsarsukite-bearing veinlets. The hybrid rock is dense and flinty even though cut by myriads of veinlets. Parts of the hybrid rock are mottled and show clearly the outlines of fragments of highly metamorphosed sedimentary rock. The breccia fragments average one centimeter in longest dimension. The parts of the hybrid rock that can be identified as syenite by their darker color are in general too fine-grained to permit recognition of minerals with a hand lens, although a few phenocrysts of feldspar about two millimeters in diameter are visible in some specimens.

The hybrid rock, as studied in thin section, is composed of porphyritic syenite enclosing vaguely defined darker and lighter fragments that differ from the syenite in proportions rather than kinds of minerals. Feldspar crystals averaging 0.5 mm. in size make up 95 per cent of the phenocrysts or about 20 per cent of the rock, and orthoclase greatly exceeds albite in amount. Aegirine phenocrysts are rare and are smaller. Many phenocrysts are broken or deformed.

The finer-grained hybrid rock is holocrystalline, with euhedral aegirine, and anhedral feldspar and quartz. Minerals in the fine-grained syenite and the breccia fragments average 0.01 mm. in size. Orthoclase and albite make up about 65 per cent of the rock, while aegirine makes up about 25 per cent. Quartz and very minor amounts of biotite and apatite are also present as well as calcite, pectolite, and unidentified cloudy alteration products of feldspar. Pectolite and calcite have replaced some of the feldspar. The pectolite is the same variety in the hybrid rock and in the veins, and is described in the section on descriptive mineralogy. Narsarsukite occurs in the hybrid rock only in tiny pegmatitic patches or veinlets.

Feldspar and Aegirine of the Hybrid Rock

The composition of the potassic feldspar is $\text{Or}_{90}(\text{Ab} + \text{An})_{10}$ as determined by the (201) method (Bowen and Tuttle, 1950). This single phase monoclinic feldspar has the optic plane normal to [010], is biaxial negative, and has a large optic angle. The properties are those of orthoclase (MacKenzie and Smith, 1956, Fig. 1).

The plagioclase is low albite of composition $An_5(Ab+Or)_{95}$ according to the indices of refraction (Chayes, 1952) and the $(131)-(1\bar{3}1)$ separation (Smith, J. R., and Yoder, 1956; Smith, J. V., 1956). The orthoclase and albite show some zonal growth, and twinning is common in albite.

Aegirine is present as euhedral rods and needles, as sheaves and radiating clusters, and as irregular clots. It is highly pleochroic, $X = \text{green}$, $Y = \text{blue green}$, and $Z = \text{yellowish green}$, and the absorption is $X = Y < Z$. The optic sign is negative, the optic angle is estimated to be 80° , the dispersion is strong, and $Z \wedge c$ is 79° degrees.

Pectolite-Quartz-Narsarsukite Veinlets

Hundreds of closely spaced veinlets make up as much as ten per cent of the rock in irregularly oriented fracture zones that cut the hybrid rock. The veinlets range from 0.01 mm. to 5 mm. in thickness, and are tens of centimeters in length. They cut phenocrysts and relict fragments alike, but apparent displacement along the fractures does not exceed a few millimeters. Pectolite is the dominant vein mineral, making up approximately half of the material in the veins. Quartz and narsarsukite are also abundant, making up 25 and 20 per cent respectively of the vein material. Calcite and galena are the remaining primary minerals in the veins. The paragenetic sequence appears to be pectolite and narsarsukite oldest, followed by galena and calcite, with quartz youngest of all.

The minerals of the veins are weathered in some specimens. The pectolite has been attacked most readily, and has darkened to a buff or brown color, or has become a cellular mass. The absence of black oxides in the alteration products suggests that the manganese content of the pectolite was low. The removal of pectolite and calcite in solution usually has been completed before the narsarsukite has been noticeably affected. Altered narsarsukite is a light buff or yellowish-brown material that probably contains leucoxene and other minerals. Galena cubes have become mantled with chalky cerussite and have persisted probably by virtue of this protective covering. The last stages of weathering have resulted in the solution of all the minerals except quartz, leaving an open meshwork of poorly shaped crystals. Specimens in this stage of weathering are common in the float in Sage Creek and other creeks draining East Butte, suggesting the narsarsukite-bearing veins may be quite common in unexposed parts of the area.

DESCRIPTIVE MINERALOGY

Narsarsukite

Narsarsukite occurs as honey yellow euhedral tablets flattened on $\{001\}$ and oriented such that the c -axis of most crystals lies in the plane of the veinlet. The largest crystals are seven millimeters across and three

millimeters thick. Some crystals are translucent and show faint symmetrical zones that differ slightly in color.

Prismatic cleavage on $\{010\}$ is distinct, and is fair on $\{110\}$. The hardness is 6 on the basal plane and in the prism zone. The specific gravity was determined with a Berman microbalance, and the average of five determinations on samples weighing from 13 to 20 milligrams was $2.783 \pm .014$.

Portions of ten crystals were examined with the optical goniometer. None of the crystals had perfect morphology, but all showed prominent

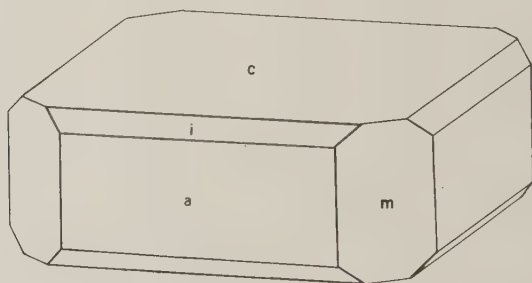


FIG. 1. Habit of narsarsukite from Sage Creek, Sweetgrass Hills, Montana.

basal faces and distinct first and second order prisms. Opposed tetragonal dipyrarnidal faces were observed on some crystals. An idealized crystal of typical habit is shown in Fig. 1. The crystals showed other forms, but their development was such that no crystal had more than one or two faces of each additional form, and no decision can be made from the morphology as to the symmetry along the two equal axes. The $\{hk0\}$ faces are very small. The morphological work by Flink (1901) and the single crystal x -ray studies by Gossner and Strunz (1932) and Warren and Amberg (1935) show that the horizontal axes are not two-fold and do not contain a vertical plane of symmetry. However, it is certain that narsarsukite crystals have a center of symmetry. The center of symmetry is indicated from the new morphological evidence, from the natural etch pits on the crystal faces, from weak symmetrical compositional zoning indicated by the color of the crystals, and by a negative test for piezoelectricity with a Giebe-Schiebe detector. The symmetry class is therefore tetragonal dipyrarnidal, $4/m$.

Crystals oriented according to the axial ratio obtained from the x -ray study described below show the following forms: $\{001\}$, $\{010\}$, $\{110\}$, $\{130\}$, $\{120\}$, $\{011\}$, $\{021\}$, $\{221\}$. The order of prominence of the forms is $\{001\}$, $\{010\}$, $\{110\}$, $\{011\}$, $\{120\}$, $\{130\}$, $\{021\}$, $\{221\}$.

TABLE 1. OPTICAL PROPERTIES OF NARSARSUKITE

	Sage Creek, Montana	Halfbreed Creek, Montana (Graham)	Greenland (Graham)
ϵ	1.658-1.662	1.653-1.655	1.625-1.636
ω	1.610-1.614	1.608-1.612	1.604-1.609
Sign	Uniaxial (+)	Uniaxial (+)	Uniaxial (+)
Birefring.	.049	.041-.046	.021-.029
Pleochroism:	E = honey yellow O = colorless	E = honey yellow O = colorless	E = colorless or pale yellow O = colorless

The optical properties of the narsarsukite from Sage Creek are cited in Table 1 together with similar data for material from other occurrences. The values obtained by Graham were verified by examination of material he collected. Narsarsukite from Sage Creek has slightly higher indices and birefringence than Graham's material.

An indexed *x*-ray powder-diffraction pattern of narsarsukite from Sage Creek is given in Table 2. The cell dimensions determined from this

TABLE 2. X-RAY POWDER-DIFFRACTION DATA, NARSARSUKITE, $\text{Na}_2\text{TiSi}_4\text{O}_{11}$,
SAGE CREEK, SWEETGRASS HILLS, MONTANA

Tetragonal dipyramidal; $I 4/m$ (C_{4h}^5), $a_0=10.72_0$ Å, $c_0=7.94_8$ Å, $c/a=0.741$.

CuK α (Ni filter), $\lambda = 1.5418 \text{ \AA}$; Camera diameter 114.59 mm.; $d_{(\text{obs.})}$ cut-off at 15.0 \AA ; corrected for film shrinkage.

<i>hkl</i>	<i>d</i> (calc.)	<i>d</i> (obs.)	I	<i>hkl</i>	<i>d</i> (calc.)	<i>d</i> (obs.)	I
110	7.580	7.609	3	521	1.931	1.931	1
200	5.360	5.365	10	440	1.895	1.896	1
211	4.105	4.107	2	512	1.858	1.861	1
002	3.974	3.976	5	600	1.787	1.787	3
220	3.790	3.787	3			1.716	1
310	3.390	3.394	8			1.695	1
301	3.259	3.260	8			1.665	2
202	3.192	3.191	2			1.630	1
222	2.743	2.745	1			1.593	1
400	2.680	2.674	1			1.550	2
312	2.579	2.579	6			1.513	1
330	2.527	2.524	6			1.488	1
420	2.397	2.396	1			1.452	1
213	2.319	2.320	2			1.410	1
402	2.222	2.221	1			1.391	1
510	2.102	2.103	1			1.369	1
422	2.053	2.054	1			1.327	3
004	1.987	1.982	3				
323	1.978			18 additional weak lines			

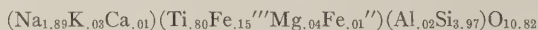
TABLE 3. CELL DIMENSIONS OF NARSARSUKITE

	Stewart (Montana)	Gossner and Strunz* (Greenland)	Warren and Amberg* (Greenland)
<i>a</i>	10.72 ₀	10.80	10.76
<i>c</i>	7.94 ₈	8.01	7.92
<i>c/a</i>	0.741 ₄	0.741	0.735

* kX converted to Å by multiplication by 1.00202 (Bragg, 1947).

pattern are listed in Table 3 and compared with values obtained by Gossner and Strunz (1932) and Warren and Amberg (1934) from specimens from Greenland. Fair agreement is shown. X-ray powder diffraction patterns of narsarsukite from Sage Creek, Montana, and Narsarsuk, Greenland, have identical spacings and intensities.

The molecular weight of the unit cell of narsarsukite from Sage Creek was calculated to be 1532 from the measured cell dimensions and specific gravity, which is equal to four formula units of weight 383. This corresponds very closely with the formula weight of $\text{Na}_2\text{TiSi}_4\text{O}_{11}$ of 382. Calculations of this sort by Warren and Amberg, and Gossner and Strunz yielded similar results. The chemical analysis by Ellestad (Graham, 1935) of narsarsukite from Montana calculates to the following when the determinations of ferric and ferrous iron by Schaller (Graham, 1935) are substituted for Ellestad's value:



The chemical evidence indicates that narsarsukite from Montana has much less substitution of ferric iron for titanium than does the material from Greenland. To check if this difference was the principal one causing the higher birefringence of the Montana material, quantitative spectrographic analyses on carefully purified material from Sage Creek and from Narsarsuk (U. S. National Museum No. 94332, $\omega = 1.608$, $\epsilon = 1.631$) were performed by Harry J. Rose, Jr., of the U. S. Geological Survey. The spectrographic analysis confirmed that iron was more abundant in the specimen from Narsarsuk, and titanium was more abundant in the specimen from Sage Creek. In addition, narsarsukite from Sage Creek was found to be significantly richer in niobium (0.6% Nb vs. 0.2%) and vanadium (0.2% V vs. 0.03%), both of which tend to increase the refraction and birefringence.

Pectolite

Pectolite occurs as white or light buff blades and clusters of radiating prisms and fibers lying in the plane of the vein. Some replacement of

feldspar by pectolite was noted along the vein borders. The crystals are elongated along the b -axis, and are flattened on $\{100\}$. The largest crystals seen were three centimeters long and three or four millimeters wide. Their thickness is less than a millimeter. The $\{100\}$ cleavage is well developed and is more perfect than a distinct cleavage on $\{001\}$. Some of the crystals are twinned on (100), and others contain many low birefringent inclusions flattened parallel to (100). The optical properties of the pectolite are as follows: $\alpha=1.603$; $\beta=1.609$; $\gamma=1.640$ (± 0.002); biaxial positive; $2V \approx 45^\circ$ (obs), 48° (calc); dispersion weak, $r < v$; $X \wedge c = 14^\circ$. Cleavage plates on (001) extinguish parallel to the trace (100); (100) plates extinguish 4° from the trace of (001). An x -ray powder-diffraction photograph was identical with one of pectolite from Bergen, New Jersey. The optical properties correspond to those of manganpectolite containing approximately 10 mol per cent $Mn_2NaSi_3O_8(OH)$ (Schaller, 1955, 1025).

Quartz, Calcite, and Galena

Quartz occurs as smoky or colorless subhedral to euhedral crystals averaging 1 to 2 millimeters in size, and only a few crystals are 3 to 4 millimeters in longest dimension. In a few cases euhedral quartz crystals which project into veinlets grew in optical continuity with broken quartz grains along the margin of the fracture. Calcite occurs as irregular white to colorless grains that rarely exceed 2 millimeters in size. Cubes of galena 1 millimeter on an edge are found associated with the calcite and quartz.

COMPARISON WITH OTHER LOCALITIES

Graham's locality near Halfbreed Creek was not visited but specimens collected by him and now in the collection of the U. S. National Museum (No. 105717) were examined. The epidote reported by Graham was not found and is believed to have been aegirine, and the clinozoisite reported is pectolite. No alunite was found.

The mineralogy at Graham's locality differs from that at Sage Creek in that the hybrid rock is somewhat richer in quartz, and the narsarsukite is not restricted to veins. Pectolite is not nearly as abundant, but much more galena is present. Paragenetic relations suggest that narsarsukite, pectolite, galena, and quartz were essentially contemporaneous.

Descriptions of the paragenetic sequence and assemblages at Narsarsuk, Greenland, are given by Flink (1901) and Gordon (1924), and are well summarized by Bøggli (1953, 319): Narsarsukite "is generally accompanied by the minerals which most frequently occur in the pegmatite veins of this locality viz., feldspar and aegirine, but it is most charac-

teristic of the mineral that it nearly always occurs together with quartz. Of the minerals mentioned the aegirine is as a rule older than narsarsukite, the others younger, which also applies to the various rarer minerals; there is however a younger generation of small crystals of narsarsukite." Microcline, albite, galena, and calcite are common members of the assemblage, but no pectolite is reported as occurring with narsarsukite. It can be seen that the assemblages are similar in Greenland and Montana.

GEOCHEMISTRY OF NARSARSUKITE

The occurrence of narsarsukite seems to depend upon circumstances where there is an excess of silica, a high partial pressure of oxygen, and an abundance of sodium and titanium. The last two conditions are probably characteristic of alkalic igneous rocks in general, but these rocks are undersaturated in silica so that a variety of unusual alkali-alkaline earth titanium silicates usually form during crystallization. It appears that reaction of emanations from alkalic magmas with siliceous wallrock is required for the formation of narsarsukite, and the formation of this mineral can be regarded in a way analogous to skarn minerals in lime-silicate metasomatism. It follows that the most probable sites for other occurrences will be where quartzites or other highly siliceous country rocks are cut by alkalic intrusive rocks and conditions are such that a high partial pressure of oxygen is present. Under conditions of lower partial pressure of oxygen, neptunite ($\text{Na}_2\text{FeTiSi}_4\text{O}_{12}$) forms, or pre-existing narsarsukite will be altered to neptunite (Flink, 1901, 129-130).

The coexistence of aegirine and narsarsukite suggests that hematite, rutile, magnetite or even ilmenite might also be expected, representing an excess of either iron or titanium. Strangely enough these minerals are not reported, and their absence is not easily explained. If saturation with iron and titanium could be established and if narsarsukite and aegirine form a discontinuous solid solution series where Fe''' substitutes for Ti in the formula $\text{Na}_2\text{Ti}_{1-x}\text{Fe}_{2x}\text{Si}_4\text{O}_{11+x}$, it might be possible to use these two minerals as a geologic thermometer. Possibly ramsayite, $\text{Na}_2\text{Ti}_2\text{Si}_2\text{O}_9$, is the sodium titanium silicate stable with quartz and rutile, and the assemblage quartz-narsarsukite-rutile is not possible. However there is as yet no evidence on this point.

The geologic evidence suggests that the total rock cover at the time of deposition of narsarsukite in the Sweetgrass Hills did not exceed 5000 feet. Assuming total vapor pressure in the mineralized zones to be equal to or only slightly greater than rock pressure, the maximum vapor pressure probably did not exceed 1000 atmospheres. The temperature of deposition cannot be determined at present, but the apparently relatively low calcium content of the hybrid rock and the presence of two

alkali feldspars in it are taken to indicate that the temperature could not have greatly exceeded 650° C. at 1000 atmospheres, otherwise only a single phase alkali feldspar would be present. The presence of pectolite is also taken to indicate relatively low intensity conditions of mineralization. The strikingly similar assemblages from Narsarsuk may also be representative of quite low intensity conditions.

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ON THE HYDRATES OF SODIUM CARBONATE, A CORRECTION, AND THE CRYSTALLOGRAPHY OF TRONA

A. PABST, *University of California, Berkeley, California.*

ABSTRACT

Reexamination of material earlier described as the hemipentahydrate of sodium carbonate, $\text{Na}_2\text{CO}_3 \cdot 2\frac{1}{2}\text{H}_2\text{O}$, proves it to be $\text{Na}_3\text{H}(\text{CO}_3)_2 \cdot 2\text{H}_2\text{O}$, trona. It is shown that all reports of the hemipentahydrate probably involved the same misidentification. A new setting of trona is based on the cell dimensions a_0 20.11 Å, b_0 3.49, c_0 10.31, β $103^\circ 8'$, space group $I2/c$. This is reconciled with previous settings and a revised angle table and indexed powder pattern are given.

Many years ago the writer, Pabst 1930, published observations supposedly applying to the hemipentahydrate of sodium carbonate, $\text{Na}_2\text{CO}_3 \cdot 2\frac{1}{2}\text{H}_2\text{O}$. This hydrate had first been reported by Morel (1889) as crystallized slowly from a solution of Na_2CO_3 between 18 and 25°C . Blasdale (1923) reported this phase as crystallizing metastably at 25°C . over a large range of composition of aqueous solutions of the carbonates and chlorides of sodium and potassium.

Mr. H. S. Peiser (1957) has kindly pointed out that the supposed hemipentahydrate was probably the sesquicarbonate, $\text{Na}_2\text{CO}_3 \cdot \text{NaHCO}_3 \cdot 2\text{H}_2\text{O}$, trona. Fortunately the material used by Pabst (1930), mostly the products of crystallizations carried out in 1926, had been preserved and reexamination quickly showed that the material earlier described as the hemipentahydrate is, in fact, the sesquicarbonate as suspected by Peiser.

Morel (1889) reported a water content of 29.7 to 30.0 per cent and Blasdale (1923, p. 2937) reported percentages of water ranging from 29.90 to 32.10. Neither specified the method of analysis. The ideal water content of the supposed hemipentahydrate would be 29.82 per cent by weight. If the sesquicarbonate is heated so that only anhydrous Na_2CO_3 remains the combined weight loss due to water and carbon dioxide driven off would be 29.66 per cent. The unreported "water content" of the material designated hemipentahydrate by Pabst (1930) was estimated from ignition loss. It may be presumed that Morel and Blasdale made their "water determinations" in the same way.

Since the "hemipentahydrate" had been described as orthorhombic by Morel (1889), a description recorded by Groth (1908, p. 196), the monoclinic sesquicarbonate must be morphologically pseudosymmetric if these materials are identical. This feature had been noted by Ayres (1889), in describing crystals of artificial "trona," who said "The symmetry of these crystals may be viewed as almost orthorhombic" (see his Fig. 2, p. 65).

The optical characterization given by Pabst (1930) was defective. The optical orientation was only partly stated as " $X=c$." Had it been fully established, Z would have been found nearly normal to a face taken as (110), in conflict with the orthorhombic interpretation. The best values for the refractive indices of trona are probably those of C. W. Bunn, reported by Brown, Peiser and Turner-Jones (1949), $\alpha=1.418$, $\beta=1.492$, $\gamma=1.543$. Of the indices reported by Pabst (1930) for the supposed hemipentahydrate, $\alpha=1.435$, $\beta=1.492$, $\gamma=1.547$, the first is clearly in error. $2V$ calculated from the indices of Bunn is $75^\circ 50'$, in close agreement with the directly determined value, $76^\circ 16'$ (yellow), of Zepharovich (1888).

The density of the supposed hemipentahydrate was given by Pabst (1930) as 2.053 on the basis of pycnometer determination. It has been redetermined on the same material by means of a Berman balance and found to be 2.11, just below the calculated density of the sesquicarbonate, 2.13, given by Brown *et al.* (1949).

In attempting to reconcile the old measurements which had been fitted to Morel's orthorhombic elements with a monoclinic interpretation it was found necessary to reexamine the several settings and elements which have been assigned to trona. Trona crystals were first measured and recognized as monoclinic by Haidinger (1825). Later the elements of Zepharovich (1888) were most widely used and they appear in the 6th and 7th editions of Dana's System of Mineralogy. They differ only slightly from those given by Ayres (1889). Earlier settings that had been chosen by Rammelsberg (1855, p. 155) and by Des Cloizeaux (1874, p. 169) found no acceptance and will not be considered here. The crystal structure of trona was described by Brown, Peiser and Turner-Jones (1949) using another setting, adopted also by Bacon and Curry (1956) and by Candlin (1956). The relations of these settings with one based on a cell having the shortest possible a and c axes are shown in Fig. 1 and Table 1. From these it will be clear that the morphological setting hitherto used corresponds to half of a centered cell, whereas doubling the a axis of this setting corresponds to choosing the shortest and next shortest translation directions in the plane of symmetry as c and a respectively, the standard monoclinic setting. The space group designation, which is $C2/c$ for the cell chosen by Brown *et al.*, then becomes $I2/c$. One may also write $I2/a$ since the space group C_{2h}^6 has two sets of glide planes.

Retabulation of the older goniometric measurements (Pabst, 1930) together with new measurements on crystals selected from the old material, on natural trona crystals from Searles Lake, California, and on newly prepared crystals of the sesquicarbonate (to be described below) as well as direct determination of the cell constants in the newly adopted

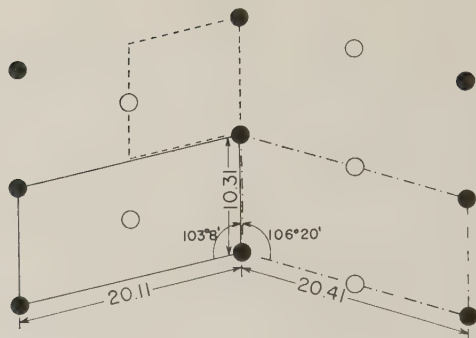


FIG. 1. The lattice of trona projected onto (010). Open circles represent lattice points displaced by $b/2$ from the level of those represented by solid circles. The C-centered cell of Brown, Peiser and Turner-Jones (1949) is shown by dot dash lines. The I-centered cell resulting from choice of the smallest possible a_0 and c_0 is shown by full lines and a cell corresponding to the axial elements of Zepharovich (1888) or of Ayres (1889) is shown by dashed lines. (See also Table 1.)

setting, all showed agreement within the limits of error with the axial elements for the new setting calculated from the cell constants given by Brown, Peiser and Turner-Jones (1949). Since the difference between these elements and those of Zepharovich involves differences in coordinate angles of more than half a degree, a new angle table has been calculated for trona, Table 2. The forms included in this table are those reported by Ayres plus $\{304\}$ (old indices), a form referred to as prominent

TABLE 1. AXIAL ELEMENTS FOR SODIUM SESQUICARBONATE DIHYDRATE, TRONA, AND TRANSFORMATIONS

	Zepharovich (1888)	Brown, Peiser and Turner-Jones (1949)	Pabst (this paper), calculated from data of Brown <i>et al.</i> (1949)
	$a:b:c$ 2.8459:1:2.9696 $\beta = 102^\circ 37'$	$a_0 = 20.41 \pm 0.03 \text{ \AA}$ $b_0 = 3.49 \pm 0.01$ $c_0 = 10.31 \pm 0.01$	$a_0 = 20.11 \pm 0.03 \text{ \AA}$ $b_0 = 3.49 \pm 0.01$ $c_0 = 10.31 \pm 0.01$
	Ayres (1889) $a:b:c$ 2.8426:1:2.9494 $\beta = 103^\circ 29'$	$a_0:b_0:c_0$ 5.848:1:2.954 (2×2.924) $\beta = 106^\circ 20'$	$a_0:b_0:c_0$ 5.763:1:2.954 (2×2.881) $\beta = 103^\circ 8'$
Zepharovich or Ayres		20I/0I0/001	2C0/010/001
Brown <i>et al.</i>	$\bar{1}0\frac{1}{2}/0\bar{1}0/001$		I0I/0I0/001
Pabst	$\frac{1}{2}00/010/001$	I0I/0I0/001	

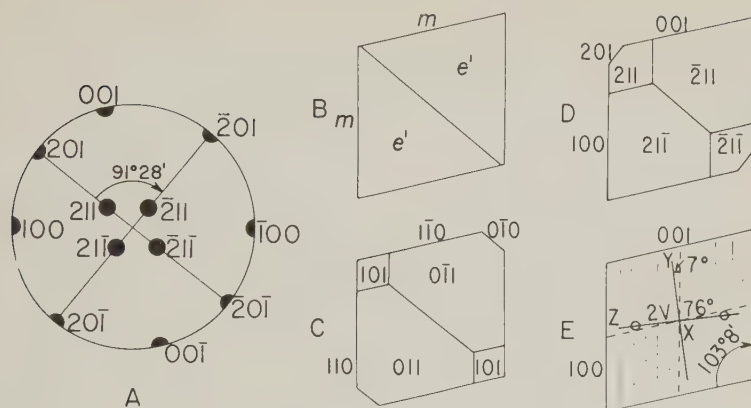


FIG. 2. A. Stereographic projection of the principal forms of trona on (010). B. End view of supposed hemipentahydrate as described by Morel (1889). C. End view of supposed hemipentahydrate as described by Pabst (1930), orthorhombic indexing. D. End view of trona as pictured by Ayres (1889). E. Cleavage and optical orientation of trona according to Zepharovich (1888) as seen in cross section normal to b .

on some crystals by Zepharovich, who also reported five other $\{h0l\}$ forms. In the course of the present study over 20 forms were identified, including a dozen $\{h0l\}$ forms not previously reported. However, the habit of both natural and artificial crystals was found to be similar to

TABLE 2. ANGLE TABLE FOR TRONA CALCULATED FOR NEWLY DERIVED ELEMENTS BASED ON X-RAY MEASUREMENTS OF BROWN, PEISER AND TURNER-JONES, (1949). COMPARE TABLE 1

Monoclinic; prismatic— $2/m$

$a:b:c = 5.763:1:2.954$ $\beta 103^\circ 8'$ $p_0:q_0:r_0 = 0.5126:2.8768:1$

$r_2p_2q_2 = 0.3476:0.1782:1$ $\mu 76^\circ 52'$ $p_0' 0.5264, q_0' 2.9540, x_0' 0.2332$

Indices			ϕ	ρ	ϕ_2	$\rho_2=B$	C	A
Dana*	New	B.P.&T.-J.						
001	001	101	90°00'	13° 8'	76°52'	90°00'	—	76°52'
100	100	100	90 00	90 00	0 00	90 00	76°52'	—
304	302	502	90 00	45 39	44 21	90 00	32 31	44 21
101	201	301	90 00	52 8	37 52	90 00	39 00	37 52
302	301	201	—90 00	53 23	143 23	90 00	66 31	143 23
111	211	311	23 32	72 45	37 52	28 53	67 58	67 35
$\bar{1}11$	211	111	—15 30	71 56	129 20	23 38	75 52	104 43
211	411	511	38 22	75 8	23 9	40 44	67 17	53 8

* Indices as given in *The System of Mineralogy*, 7th ed., vol. II, p. 138, 1951 referred to the elements of Zepharovich (or Ayres). Compare Table 1.

TABLE 3. REINTERPRETATION OF INTERFACIAL ANGLES REPORTED BY PABST (1930) FOR THE SUPPOSED HEMIPENTAHYDRATE

Morel (Pabst, 1930)	Measured	Calculated	Zepharovich	Pabst	Calculated
101 \wedge 101	57°57'	57°52'	111 \wedge 111	211 \wedge 211	57°46'
011 \wedge 011	47 44	*47 24	111 \wedge 111	211 \wedge 211	47 16
110 \wedge 110	76 27	†76 54	001 \wedge 100	001 \wedge 100	76 52
010 \wedge 011	66 17	66 18	101 \wedge 111	201 \wedge 211	66 22
010 \wedge 110	51 46	51 33	101 \wedge 001	201 \wedge 001	52 28
010 \wedge 110	51 47	51 33	101 \wedge 100	201 \wedge 100	50 40
					51 34

* Should have been given as 47°26' to correspond to Morel's measured value. Discrepancy is due to calculation from rounded value of c/b , 0.439.

† This angle is erroneously given by Groth (1908) as 71°54' due to the misprint in Morel's report commented upon by Pabst (1930, p. 70).

the habits described by Zepharovich and Ayres except for the constant presence of the form $\{201\}$ (new indices) on the artificial crystals grown in solutions containing potassium, as shown in Fig. 2C.

It is now possible to reinterpret the angle measurements that had been fitted to the orthorhombic elements of Morel. In Table 3 the old measurements (Pabst, 1930, Table 1) are reproduced together with the revised interpretation. The first four angles listed do not differ by more than 8' in the calculated values for the two settings. The error of the orthorhombic setting arises from the failure to recognize that the last two angles are not equivalent. The average of the ideal values of these angles matches closely the ideal value in the faulty interpretation. The pseudoorthorhombic cell obtained from the cell of Brown *et al.* by the transformation $101/\overline{103}/040$ or from the newly chosen cell by the transformation $102/10\overline{2}/040$, has the axial elements $a:b:c=0.793:1:0.437$, $\alpha=\beta=90^\circ$, $\gamma=91^\circ 28'$, very close to the orthorhombic elements given by Morel, $a:b:c=0.794:1:0.439$, $\alpha=\beta=\gamma=90^\circ$. Evidently Morel made very good measurements.

Figure 2 shows the pseudoorthorhombic character of trona in stereographic projection, together with several (010) elevations and a diagram of the optical orientation as determined by Zepharovich (1889, p. 138). All other published observations on the orientation of trona are essentially in agreement with this.*

It is quite certain that the crystals described by Pabst (1930) as the hemipentahydrate of sodium carbonate and now shown to be the sesquicarbonate were the same as those described as the hemipentahydrate by

* The statement in Winchell's Elements of Optical Mineralogy, pt. II, 4th ed., 1951, page 126 that " $Y \wedge c = -17^\circ$ " clearly should be " $Y \wedge c = -7^\circ$."

Blasdale (1923). They were obtained in the course of evaporation of an equimolal solution of sodium and potassium carbonates at 25° C., that is, within the range of compositions in which Blasdale obtained his "hemipentahydrate." Apparently the excess of CO₂ in the crystals entered the solution from the air since only neutral carbonates were used as starting materials.

The experiment of Morel was imitated by slowly evaporating solutions of Na₂CO₃ at room temperature, 24 to 28° C., slightly warmer than the temperature range of the original. Slow evaporation in crystallizing dishes protected only by a dust cover and with free access of air yielded clusters of the acicular sesquicarbonate crystals together with more or less equant crystals of the monohydrate (thermonatrite). In one series of crystallizations the crystals formed after a few days were redissolved by addition of a little water. This was repeated and upon the third and final evaporation of this lot abundant sesquicarbonate formed before any monohydrate.

Several studies, summarized by Seidell (1940, pages 1193-1198), have shown that the sesquicarbonate may be precipitated at very low concentrations of bicarbonate. At room temperature, under equilibrium, at extremely low bicarbonate concentrations, it may be accompanied by the decahydrate or possibly the heptahydrate. In only one trial was one of these, probably the decahydrate, obtained. It formed alone in clear, platy crystals by rapid precipitation from a solution of small volume and large surface area. The crystals effloresced rapidly in dry air, being transformed into pseudomorphs by coalescence of many clusters of the dehydration product arising at various points in each crystal. The process can be easily followed with a binocular microscope, and the pseudomorph of a suitably small, elongate, crystal makes a perfect powder preparation yielding an x-ray diffraction pattern showing only the lines of the monohydrate, thermonatrite.

The ASTM X-ray powder data file contains 3 cards for trona, 1-0938, 1-1077 and 2-0601. The first of these is stated to be for the decahydrate, though the material is named "trona" and the optical properties and density of trona are given on the card. The pattern recorded on card 1-0938 however, is not that of trona and presumably is that of the decahydrate. It coincides with the decahydrate pattern given under number 794 by Hanawalt, Rinn and Frevel (1938). Pattern number 793 of these authors, referred to Na₂CO₃ · 2½H₂O, is the sesquicarbonate pattern. This was taken into the ASTM file as 1-1077 for trona implying a correction analogous to that here set forth. Table 4 records a powder pattern for trona indexed for all lines to 2.0 Å. The indexing has been checked by comparison with Weissenberg and precession patterns.

The false reports of the existence of Na₂CO₃ · 2½H₂O have given rise to

TABLE 4. X-RAY POWDER DIFFRACTION PATTERN OF SODIUM SESQUICARBONATE, TRONA. Cu RADIATION, Ni FILTER

Observations on trona from Searles Lake, California, and artificial material agree exactly. Calculated spacings for all possible lines with $d > 2.00$ Å, for cell of dimensions a_0 20.11 Å, b_0 3.49, c_0 10.31, β 103°8'; space group $I2/c$

hkl	$I_{\text{obs.}}$	$d_{\text{obs.}}$	$d_{\text{calc.}}$	hkl	$I_{\text{obs.}}$	$d_{\text{obs.}}$	$d_{\text{calc.}}$
200	ms	9.88 Å	9.79 Å	404			2.475
002	—	—	5.02	512			2.471
202			4.95	800	ms	2.447	2.448
	m	4.92					
400			4.90	213			2.441
202	vw	4.12	4.11	802	w	2.426	2.429
402	w	4.00	3.99	013			2.417
110	w	3.43	3.44	611			2.411
011			3.30	413			2.323
600			3.26	204			2.310
	ms	3.21					
211			3.20	213	m	2.259	2.262
402			3.17	611			2.238
310			3.08	710			2.183
	s	3.08			vw	2.186	
602			3.07	512			2.182
211			3.06	712	w	2.149	2.146
112			2.89	613	w	2.119	2.115
411			2.83	114			2.067
					w	2.060	
112	vw	2.79	2.79	314			2.057
312	mw	2.76	2.76	404			2.053
411	vs	2.659	2.645	811	m	2.040	2.039
510			2.606	413			2.038
					m	2.032	
204	w	2.587	2.578	802			2.027
				—			—
004			2.512		mw	1.996	
	mw	2.510					
312			2.508		mw	1.965	
602	vw	2.485	2.491		w	1.886	

and over 30 additional lines

confusing statements in several standard works of reference. In Winchells' *Microscopic Characters of Artificial Inorganic Substances*, 2nd ed., New York, 1931, the hemipentahydrate is described on page 200 largely on the basis of Pabst's faulty work and in a footnote it is suggested that "Larsen's data for 'trona' from Vesuvius may apply to this substance," a most unfortunate reversal of the situation as it is now revealed. In *The Barker Index of Crystals*, Volume 1, Part 2, Cambridge, 1951, the hemipentahydrate is described under number 0.369. This description is

largely based on Groth, but, the misprint of an interfacial angle taken over by Groth from Morel (see second footnote to table 3 of this paper) not being recognized, new axial elements are calculated and a "correction" of Groth's elements which had been properly copied from Morel is given. These new elements correspond to no crystallographic measurements at all, not even to an orthorhombic interpretation of trona.

Equilibrium crystallization of $\text{Na}_2\text{CO}_3 \cdot 2\frac{1}{2}\text{H}_2\text{O}$ over a wide range of compositions at 0° and 15° C. in the system $\text{H}_2\text{O}-\text{NH}_4\text{Cl}-\text{Na}_2\text{CO}_3$ was reported by Mondain-Monval (1922) who referred to this phase as "carbonate de Morel." His work is the basis of Fig. 31, page 383, in the *International Critical Tables*, vol. IV, New York, 1928. Mondain-Monval did not report how he identified the phase but it may be confidently assumed that this was another case of failure to recognize the appearance of the sesquicarbonate.

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THE ORIGIN OF THE VERMICULITE DEPOSIT AT LIBBY, MONTANA

WILLIAM A. BASSETT, *Columbia University, New York, N. Y.*

ABSTRACT

The country's largest vermiculite mine at Libby, Montana has been studied and the origin of the vermiculite deposit investigated. The ore body lies in an augite pyroxenite, which has been altered to biotite, hydrobiotite, and vermiculite. Numerous syenite dikes may be observed in the pyroxenite.

Ion exchange experiments show that biotite alters to vermiculate at room temperature in solutions of 0.001 molar magnesium or calcium chloride. Roy and Komo's evidence for the instability of vermiculite under hydrothermal conditions and its application to the origin of the vermiculite deposit at Libby, Montana is discussed. A mechanism is suggested for the formation of hydrobiotite, a regularly interstratified biotite-vermiculite.

It is concluded that augite altered to biotite at the time of the intrusion of the syenite dikes, and that the biotite was altered to hydrobiotite and vermiculite by supergene solutions.

INTRODUCTION

Vermiculite deposits of non-sedimentary origin are widely recognized, and many of them are similar in certain aspects. Vermiculite occurs as an alteration product of mafic minerals in an ultramafic body which has been intruded by small dikes and sills of acidic rock, usually pegmatites. The vermiculite usually occurs in fractures, along contacts of the ultramafic body with the country rock and along contacts between the ultramafic rock and the acid rock. It is accompanied by other alteration minerals such as antigorite, chrysotile, chlorite, talc, pyrophyllite, biotite, and amphiboles.

Vermiculite occurrences have been attributed both to hydrothermal and to supergene origins by different workers. Kulp (1954), in his study of the Day Book Dunite of Yancey County, North Carolina, attributed vermiculite to the supergene alteration of phlogopite. Hadley (1949) concluded that the vermiculite at Buck Creek, Clay County, North Carolina was the direct consequence of the intrusion of the pegmatite. Hagner (1944) studied the Wyoming vermiculite deposits and found evidence for the hydrothermal origin of the vermiculite. Thus, the origin of vermiculite deposits formed in seemingly similar environments has received different interpretations.

Libby, Montana is petrologically and mineralogically one of the simplest of the vermiculite deposits. Three sheet silicate minerals, biotite, hydrobiotite, and vermiculite, are involved. The original ultramafic was nearly a single mineral, augite, and the intruding acid rock, a syenite, is of simple composition. The simplicity of the mineral relationship has made it possible to study each chemical change and mechanism involved in the formation of the vermiculite deposit.

The term *vermiculite* in this paper refers to the mineral as defined by x-ray, optical, and chemical properties (Grim, 1953). The terms *vermiculite deposit* and "*vermiculite*" refer to the material of commercial interest which generally contains biotite and hydrobiotite in addition to vermiculite.

THE MINING OPERATION

The Zonolite Company operation, about seven miles northeast of Libby, Montana, is at present the largest vermiculite mine in the country. The mine is on a hill about 4,000 feet above sea level and 2,000 feet above the Kootenai River, which flows through the valley three miles from the mine.

The mine is a large open pit (Fig. 1) with ten benches and has been in operation since 1923. The ore is handled by power shovels and heavy trucks. Little blasting is necessary, since the ore is soft and can be worked directly by the shovels. One ton of waste is mined for each ton of ore, and approximately 1,200,000 tons of ore are processed each year in a mill located just below the mine. The mill produces about 140,000 tons of concentrate each year, employing both wet and dry processes. This concentrate is shipped to various distributing points throughout the country, where it is heated in furnaces to 2,000° F., causing expansion to as much as 20 times its original volume. In this form it is marketed for insulation and horticultural purposes. The early history of the operation is described by Kriegel (1940).

GEOLOGICAL SETTING

The ore body is a pyroxenite which intruded the Belt Series sediments and was later altered. Pardee and Larsen (1929) describe the Belt Series



FIG. 1. The upper benches of the Zonolite mine at Libby, Montana.

in this area as argillites and quartzites which exhibit moderately steep dips and open folds that trend northwesterly. Near the borders of the pyroxenite, the argillites show metamorphism to hornstone which in places carries magnetite.

Pardee and Larsen's geologic map shows the pyroxenite as an irregularly shaped body about 3.5 miles long by 1.5 miles wide, with its long axis north-south. The original material was almost entirely augite, which was invaded by syenite dikes and subsequently altered to biotite, hydrobiotite, and vermiculite.

Southwest and adjacent to the pyroxenite is a syenite intrusive about two-thirds as large, which consists principally of potash and soda feldspars. Pardee and Larsen (1929) report hornblende, fluorite, apatite, sphene, rutile, biotite, and garnet as accessory and secondary minerals. A dike-like projection of the syenite extends into the pyroxenite. The syenite dikes which cut the pyroxenite range from a few inches to several feet in thickness (Fig. 2) and were probably fed by the major projection of the syenite.

Pyroxenite

The pyroxenite was probably originally an almost pure coarse-grained augite. Crystals as much as 3 inches long are easily discernible, even



FIG. 2. A typical syenite dike cutting the altered pyroxenite.



FIG. 3. Horizontally oriented crystals in the wall of the mine. The vertical marks at the left were made by a power shovel.

though some of them are almost completely altered. The larger crystals show a strong horizontal orientation (Fig. 3). At many points, domains of horizontally oriented crystals are distinctly cut by domains of finer grained augite showing vertical flow features. These domains of vertical flow are particularly conspicuous because of their vertical banding (Fig. 4).

New magma from below might have produced the vertical flow features, but such a mechanism should produce grosser and more continuous features rather than the small domains observed.

The contacts of the pyroxenite are characterized by a high concentration of magnetite and apatite. Where the pyroxenite is in contact with the Belt sediments, augite, magnetite, and apatite have penetrated a short distance into the metamorphosed sediments as dark fine-grained (1 mm.) bands. In the pyroxenite, magnetite and apatite are more abundant near the contact. Both minerals are fine-grained (5 mm.) and are disseminated in a relatively fine-grained augite. The apatite occurs as well crystallized hexagonal prisms. Since the rock is friable because of the alteration of the augite, the prisms readily separate. The magnetite-apatite zone appears both along the pyroxenite-syenite contact and the pyroxenite-sediment contact.

Syenite

The mine workings have exposed numerous thin syenite dikes, ranging from a few inches to several feet thick, which cut the pyroxenite (Fig. 2). The syenite consists of potash and soda feldspars intergrown as perthites in some places and discrete in others. The ratio of soda to potash feldspar varies from point to point. The individual dikes almost cer-

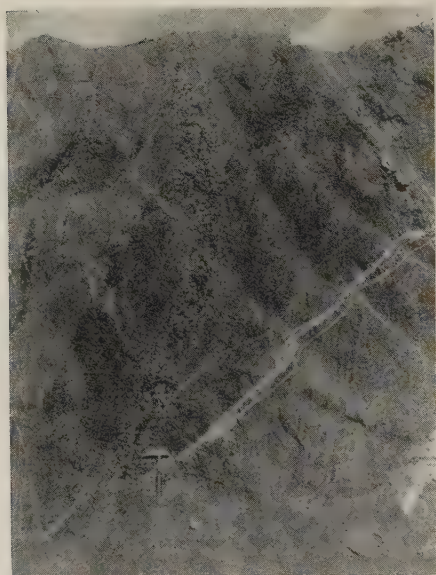


FIG. 4. Vertical flow features in the pyroxenite. These probably represent later injections of augite magma. Thin asbestos veins cross diagonally.

tainly represent different introductions of magma separated by short intervals of time. Pardee and Larsen observed minor textural variations in the main body of syenite, which they attributed to separate introductions. In some places, the syenite is quite dark, due to the presence of finely disseminated hornblende crystals. The hornblende is probably not a product of the syenite magma but is the result of assimilation of augite by the syenite magma. Figure 5 shows crystals cutting across a syenite-pyroxenite contact. These crystals are augite at the end which extends into the pyroxenite and hornblende pseudomorphous after augite at the end which projects into the syenite.

Alteration Products of the Pyroxenite

Four alteration minerals predominate, asbestos (tremolite-actinolite), biotite, hydrobiotite, and vermiculite. The name hydrobiotite was applied by Gruner (1934) to interstratified biotite-vermiculite from Libby. This mineral, along with vermiculite and biotite, constitutes the commercial "vermiculite" ore.

Many thin (approximately 1 inch), white asbestos veins cut through the pyroxenite. The asbestos has been identified by x-ray diffraction and optically as tremolite-actinolite. It differs from augite in that it contains water and has a higher silica content. Some thick asbestos veins contain

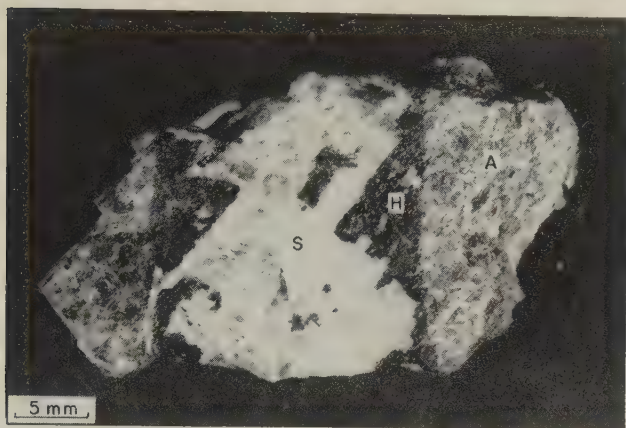


FIG. 5. A small syenite dike (S) which has altered projecting augite crystals (A) to hornblende (H).

cores of quartz, which indicates that silica-rich solutions were probably responsible for altering augite to tremolite-actinolite (Fig. 6). This quartz is the only free silica found in the mine.

The asbestos is also found disseminated through the intrusive as thin layers along cleavage planes of the augite. These tend to greatly exaggerate the cleavage, and as a result, complete crystals or even large fragments of augite are almost impossible to find.

Biotite, hydrobiotite, and vermiculite are widely distributed general alteration products of the augite. These minerals are also found along cleavage planes of the augite (Fig. 7). Their orientation is derived from the augite by this mechanism and remains even where no original augite exists.

Unaltered patches of augite several feet across are found in the mine. These appear to be domains which were not permeated by the altering solutions. Similarly, there are patches of biotite which have not been altered to hydrobiotite and vermiculite. This feature is the strongest field evidence for the derivation of hydrobiotite and vermiculite from biotite rather than from augite by hydrothermal action. The unaltered biotite patches are quite separate from the augite patches, suggesting that the alteration of augite to biotite and the alteration of biotite to hydrobiotite and vermiculite occurred at different times.

Vermiculite differs from biotite in the composition of the intersilicate layer. The intersilicate layer in biotite consists of potassium ions, while the intersilicate layer in vermiculite consists of bivalent ions distributed between two layers of water molecules. Alteration from biotite to vermiculite takes place when hydrated bivalent ions in solution replace the

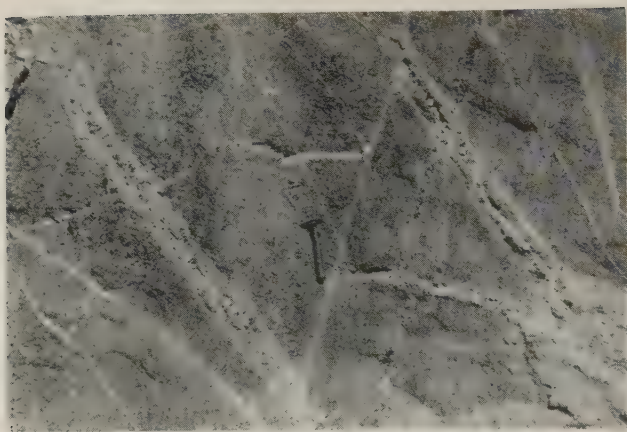


FIG. 6. An offset asbestos vein showing a core of quartz.

potassium ions in biotite. In hydrobiotite, potassium intersilicate layers alternate with intersilicate layers containing bivalent ions and water molecules. An analysis by the Zonolite Company of the exchangeable cations reveals that the concentrate from Libby contains 20.8 milli-equivalents of calcium and 12.0 milli-equivalents of magnesium per hundred grams of concentrate. A complete analysis of the concentrate is given in Table 1 (Zonolite 1954).

An oriented sample of concentrate from the mill was run on the x-ray diffractometer (Fig. 8b). The resulting pattern indicates that biotite, hydrobiotite, and vermiculite are present in about equal amounts. Only about 1 per cent of the flakes in the concentrate, however, are discrete biotite and fail to exfoliate on ignition. In the commercial process, these unexpanded flakes are separated from the final product by air sorting.

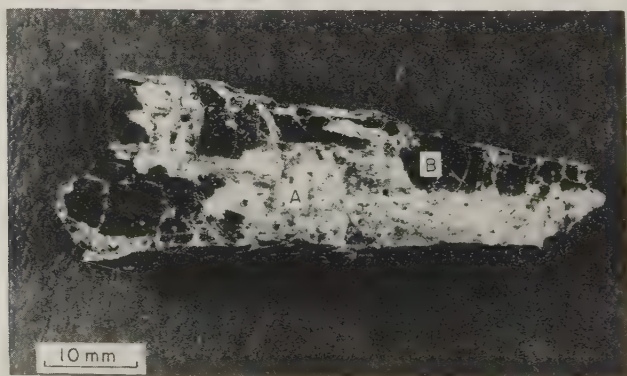


FIG. 7. Augite (A) partially altered to biotite (B) along cleavage planes.

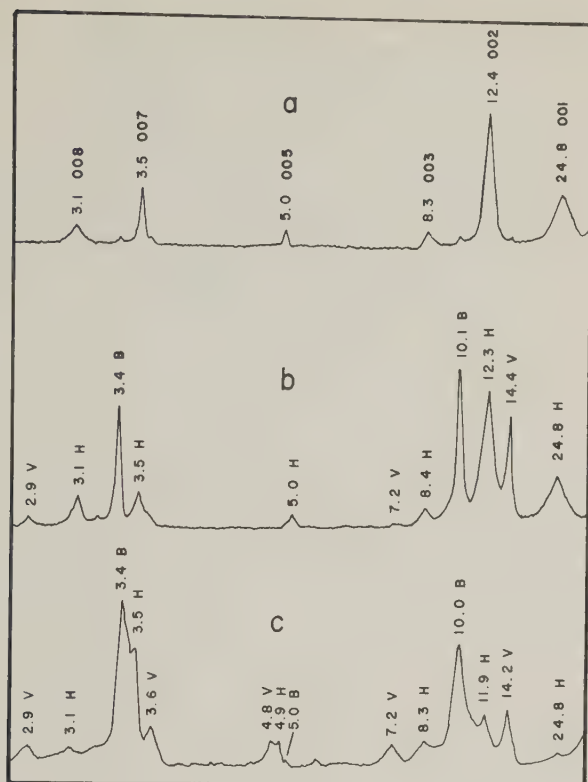


FIG. 8. X-ray diffraction data. (a) Hydrobiotite from the Zonolite mine at Libby, Montana. (b) Concentrate from the mill at Libby, Montana containing biotite (B), hydrobiotite (H), and vermiculite (V). (c) Copper hydrobiotite (H) and copper vermiculite (V) synthesized from biotite (B).

The biotite shown in the x-ray pattern is probably intimately mixed with the hydrobiotite and vermiculite.

Only hydrobiotite and vermiculite provide the water which is turned to steam in the exfoliation process. One might expect the biotite to act only as a diluent decreasing the yield of the expanded material. This is not the case, however, for the biotite compensates for its deficiency as a source of steam by providing strength to the layers and impeding the escape of steam produced by the hydrobiotite and vermiculite. The biotite interstratified with vermiculite in the hydrobiotite probably serves the same function.

A pure vermiculite from Corundum Hill, North Carolina, when ignited, shows a much greater tendency to decrepitate. Expanded concentrate from Libby placed in water and prepared for x-ray diffraction in

TABLE 1. CHEMICAL ANALYSIS OF CONCENTRATE FROM LIBBY (ZONOLITE, 1954)

SiO ₂	38.64
MgO.....	22.68
Al ₂ O ₃	14.94
Fe ₂ O ₃	9.29
K ₂ O.....	7.84
CaO.....	1.23
Cr ₂ O ₃	0.29
Mn ₂ O ₄	0.11
P ₂ O ₅	trace
S.....	trace
Cl.....	0.28
H ₂ O.....	5.29
Total.....	100.59

a Waring Blendor yields a biotite diffraction pattern. Expanded vermiculite from Corundum Hill prepared in the same way gives a diffraction pattern showing a 9.4 Å reflection resulting from the collapse of the vermiculite lattice after the interlayer water is driven out. It is of interest to note that biotite seems to be an essential constituent of commercial "vermiculite."

OPTICAL DATA

Both color and refractive index change as biotite alters to hydrobiotite and vermiculite (Table 2). The color of the unaltered biotite is gray-green, whereas the color of the hydrobiotite is light brown and vermiculite is neutral gray. These color differences are used by the miners in selecting ore and avoiding the domains of unaltered biotite. The refractive index and the birefringence decrease as the water content increases from biotite to hydrobiotite to vermiculite. The optic sign and the 2V remain unchanged.

The partial alteration of a flake of biotite to vermiculite is shown in

TABLE 2. OPTICAL DATA ON THE ORE MINERALS FROM LIBBY

Mineral	Color	α	γ	$\gamma - \alpha$	Sign	2V
Biotite	Gray-green	1.570	1.610	0.040	(-)	0-5°
Hydrobiotite	Light brown	1.560	1.595	0.035	(-)	0-5°
Vermiculite formed synthetically from biotite	Neutral gray. Berlin blue under crossed nicols	1.530	1.555	0.025	(-)	0-5°

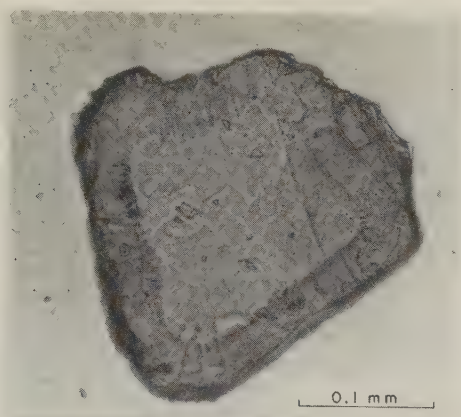


FIG. 9. A biotite flake which has been altered to vermiculite along the edges by immersion in a molar magnesium chloride solution at 100° C. for 6 hours.

Fig. 9. The biotite was placed in molar magnesium chloride at 100° C. for 6 hours. The change in refractive index shows clearly the penetration of the alteration into the flake. Under crossed nicols, the rim of vermiculite has an anomalous blue birefringence of the first order, similar to that found in penninite.

LABORATORY INVESTIGATIONS

Field observations at Libby, Montana suggest that hydrobiotite and vermiculite formed as alteration products of biotite which had previously been formed by the alteration of augite. Separate domains of unaltered biotite and unaltered augite suggest that there were two periods of alteration: soaking of the pyroxenite with alkali-silica rich hydrothermal solutions to produce biotite from the augite, and soaking of the pyroxenite, now partially altered to biotite, with magnesium-calcium rich solutions to produce hydrobiotite and vermiculite from the biotite. The depth to which the ore extends indicates that both of the altering solutions permeated the pyroxenite. Such extensive permeation suggests hydrothermal solutions; yet the high relief of the region could also permit deep penetration of the supergene solutions.

These field observations suggest some questions which might be resolved by laboratory experiments:

1. Under what conditions does biotite alter to vermiculite?
2. Is vermiculite stable under hydrothermal conditions?
3. What is the nature of the mixed layer structure observed in hydrobiotite and what conditions lead to its formation?

Alteration of Biotite to Vermiculite

Barshad (1948) conducted ion exchange experiments on vermiculite and biotite. He showed that vermiculite can be produced by placing biotite in a solution of neutral molar magnesium chloride at 70° C. for 10 days. The use of large samples (0.5–2.0 grams) made it necessary to change the solutions and wash the samples daily. He showed also that the vermiculite can be altered to biotite by placing it in a potassium solution.

The writer was able to confirm these results and extended the ion exchange experiments to include a greater variety of concentrations, temperatures, and reaction times (Table 3). The results indicate that the reaction proceeds under much less intense conditions than those used by Barshad. Some of the unaltered biotite from Libby altered partially to vermiculite in a 0.001 molar magnesium chloride solution in 56 hours at room temperature. It showed similar behavior in a calcium chloride solution of the same concentration.

The failure of the reaction to go to completion may be attributed to two conditions: (a) the inhibiting effect of potassium in the solution and (b) an inherent reluctance of some of the biotite to alter to vermiculite. Potassium is released into the solution from the biotite as the magnesium enters the biotite to replace the potassium. As the concentration of the potassium in the solution builds up, the alteration of the biotite becomes slower until equilibrium is reached and the reaction is halted. Under laboratory conditions, this point is characterized by the co-existence of discrete biotite and vermiculite. Under natural conditions of more dilute solutions, greater reaction times, and more dynamic conditions, the failure of the reaction to go to completion is believed to result in mixed-layer biotite-vermiculite. This is discussed in more detail later.

When 1 mg. of biotite from Libby was immersed in 50 ml. of 0.001 molar magnesium chloride, the biotite altered completely to vermiculite, while 50 mg. of biotite under similar treatment (Table 3) altered only partially to vermiculite. This confirms the importance of the inhibiting effect of the potassium released from the biotite.

The inhibiting effect of potassium is more pronounced when high concentrations are used. A molar solution of magnesium chloride containing 0.04 molar potassium chloride does not alter biotite to vermiculite. In more dilute solutions (0.001 molar), the potassium-magnesium ratio may be as high as 1:1 and still effect a partial alteration of biotite to vermiculite.

Two significant facts are brought out by the experiments described

TABLE 3. THE ARTIFICIAL ALTERATION OF BIOTITE TO VERMICULITE UNDER VARIOUS CONDITIONS

Sample: Unaltered biotite from Libby, Montana

Quantity of Sample: 50 mg.

Sample Size: Smaller than 149 microns.

Quantity of Solution: 50 ml.

Composition of Solution		Temperature (degrees C.)	Number of Hours	Percentage Bio- tite Alteration to Vermiculite
0.2	molar $MgCl_2$	100	56	100
0.1	molar $MgCl_2$	100	56	95
0.01	molar $MgCl_2$	100	56	60
0.001	molar $MgCl_2$	100	56	35
1	molar $MgCl_2$	R.T.*	72	50
0.2	molar $MgCl_2$	R.T.	56	35
0.1	molar $MgCl_2$	R.T.	56	35
0.01	molar $MgCl_2$	R.T.	56	25
0.001	molar $MgCl_2$	R.T.	56	15
0.01	molar $CaCl_2$	100	24	45
0.001	molar $CaCl_2$	100	24	35
0.01	molar $CaCl_2$	R.T.	24	20
0.001	molar $CaCl_2$	R.T.	24	15
0.001	molar $CaCl_2$			
0.001	molar $MgCl_2$	100	24	40
0.001	molar $CaCl_2$			
0.001	molar $MgCl_2$	R.T.	48	15
1	molar $MgCl_2$			
0.02	molar KCl	100	48	30
1	molar $MgCl_2$			
0.03	molar KCl	100	48	10
1	molar $MgCl_2$			
0.04	molar KCl	100	48	<1
0.001	molar $MgCl_2$			
0.0001	molar KCl	100	48	70
0.001	molar Mg			
0.001	molar KCl	100	48	40

* R.T. = room temperature.

above. The first is that biotite readily alters to vermiculite under conditions of very low concentrations, low temperatures, and short reaction times. The other is that when the potassium concentration in a solution exceeds 0.04 molar, biotite does not alter to vermiculite even when the magnesium concentration is molar and the temperature is 100° C.

The first of these facts shows that vermiculite can form from biotite

under supergene conditions. The second indicates that the solutions which produced the vermiculite were probably quite separate in time from the alkaline-rich solutions which must have accompanied the introduction of the syenite at Libby.

The Stability of Vermiculite under Hydrothermal Conditions

Roy and Romo (1957) subjected vermiculite to hydrothermal conditions in a bomb. They found that at 10,000 p.s.i. the vermiculite undergoes a change when it is heated above 200° C. Above 200° C., the 4.7 Å and 7.0 Å reflections on the x-ray diffraction patterns increase in intensity with respect to the 14 Å reflection. They suggest that a type of chlorite is formed, although the intensity relationships do not shift all the way to a chlorite pattern, but give a pattern representing some sort of mixture of vermiculite and chlorite. They suggest that magnesium ions migrate from the octahedral position of the vermiculite to the intersilicate position where they increase the scattering power of the intersilicate layer and decrease the scattering power of the octahedral layer to produce the changes observed in the x-ray pattern.

Roy and Romo show the DTA patterns of vermiculite before hydrothermal treatment and after hydrothermal treatment. The pattern of the vermiculite before the treatment has a pair of endothermic peaks in the range 100° C. to 300° C. The DTA pattern of the material after hydrothermal treatment shows a minor endothermic reaction at about 180° C. and a strong endothermic reaction at about 640° C.

Characteristically, vermiculite shows a pair of endothermic reactions between 100° C. and 300° C., due to the loss of H₂O from the intersilicate position. The low temperature peak is attributed to the loss of water molecules not in contact with the magnesium ions. The higher temperature endothermic peak results from the removal of the more tightly bound water molecules in contact with the magnesium ions (Mackenzie 1957, p. 194). Chlorite gives a DTA pattern with a strong endothermic reaction in the range 450° C. to 650° C. due to dehydroxylation of the brucite layer (Mackenzie 1957, p. 215).

The writer suggests a different interpretation from that offered by Roy and Romo for the changes taking place in the vermiculite when treated hydrothermally. The DTA diagrams indicate the presence of hydroxyl ions in the intersilicate position along with H₂O molecules rather than just H₂O molecules as shown in Roy and Romo's diagrams. The writer suggests that hydrolysis takes place in the bomb, providing the magnesium ions with hydroxyl ions in their hydration envelopes. Thus, more magnesium ions can be accommodated in the intersilicate position because some of the positive charge of the magnesium ions is balanced by

the hydroxyl ions. It would be interesting to know if vermiculite would alter to chlorite completely if more magnesium were made available.

Regardless of the interpretation, the data strongly suggest that vermiculite does not continue to *exist* as vermiculite under hydrothermal conditions (10,000 p.s.i. over 200° C.) and that it therefore could not *form* under hydrothermal conditions. The only reservation which the writer has in arriving at such a conclusion is that the acidity of the hydrothermal solution might have a pronounced effect on the temperature at which the transition from vermiculite to chlorite takes place if the transition is actually effected by hydrolysis. There is a need for more extensive experiments in this realm of investigation.

Hydrobiotite (mixed-layer biotite-vermiculite)

Gruner (1934) described a specimen from Libby, Montana which he found to be mixed-layer biotite-vermiculite and which he designated hydrobiotite, since it retains many of the characteristics of biotite and yet contains a large quantity of water.

Two types of mixed-layer structure exist, randomly interstratified and regularly interstratified. The randomly interstratified mixed-layer structure gives an irrational sequence of basal reflections on a diffractometer diagram. That is, the reflections cannot be indexed (001), (002), (003), etc. Instead, they occur between adjacent reflections of the component minerals. On the other hand, the regularly interstratified mixed-layer structure gives a rational sequence of reflections with a (001) reflection having a spacing which is the sum of the spacings of the components. The reflection for the sum of the spacings of the components may sometimes be observed in randomly interstratified mixed-layer structures (Weaver 1956, p. 217) but is generally broad.

Hydrobiotite from Libby, Montana was prepared in distilled water in a Waring Blendor for 5 minutes. The suspension was allowed to sit for a minute while the coarse fragments settled out. Then the fines (less than 50 microns) were transferred in suspension to a settling dish where they were sedimented onto a glass slide for 10 minutes. The water was then removed and the sample air dried.

Three samples of hydrobiotite from different parts of the mine were prepared in this way and run on the x-ray diffractometer immediately after preparation. The resulting patterns had rather broad reflections, with the most intense reflection at 11.5 Å, which were found to be in fair agreement with the values reported by Gruner (1934). The same samples, still mounted on the glass slides when rerun a month later, gave results (Fig. 8a) quite different from those obtained earlier. The reflections were sharp and the most intense reflection had shifted to 12.4 Å.

The change which took place is attributed to a change in the degree of hydration of the vermiculite layers in the hydrobiotite. The vermiculite layers in the original sample are probably partially and randomly hydrated. After grinding to a fine particle size and exposure to the air for a period of a month, the layers picked up water and expanded to the full 14 Å.

A fully hydrated sample gives a rational series of basal reflections indicating that it is regularly interstratified (Fig. 8a). The spacing represented by the (001) reflection is almost exactly the sum of the (001) spacings for the biotite (10 Å) and the vermiculite (14 Å) found at Libby, indicating that the ratio of biotite to vermiculite is 1:1 and almost certainly a sequence of BVBVBVBVBV.

The hydrobiotite occurs in the mine as domains several feet across in which there is almost no discrete biotite and vermiculite, as indicated by the diffraction pattern of hydrobiotite collected *in situ* (Fig. 8a). A sample of concentrate from the mill served as a representative blended sample of the micaceous minerals from a large portion of the mine. A diffraction pattern of this material (Fig. 8b) shows that hydrobiotite occurs in the same order of abundance as biotite and vermiculite in the mine. The sharpness of the hydrobiotite reflections indicates that there is no appreciable variation in the biotite to vermiculite ratio in the hydrobiotite, even when material is sampled from a large portion of the mine. The wide distribution of the 1:1 regularly interstratified biotite and vermiculite seems to indicate that this is a relatively stable configuration.

None of the ion exchange experiments involving the immersion of biotite in magnesium and calcium solutions yielded hydrobiotite. However, mixed-layer biotite-vermiculite has been produced experimentally in two ways: (a) by immersing 10 mg. of biotite in molar cupric chloride for 100 hours at 100° C. (Fig. 8c)* and (b) by immersing vermiculite in solutions of varying potassium concentrations. The former of these methods yielded a regularly interstratified 1:1 mixed-layer structure and has been designated copper hydrobiotite (Bassett, 1958).

The positions and intensities of the copper vermiculite reflections are different from those of magnesium and calcium vermiculite because hydroxyl ions accompany the copper in the intersilicate position. The latter method produced mixed-layer structures of varying ratios, depending on the concentration of the potassium, the temperature, and the reaction time. It showed no tendency to form a stable ratio as found in the natural hydrobiotite or the synthesized copper hydrobiotite. This suggests that the hydrobiotite found at Libby, Montana formed from the action of dilute magnesium and calcium solutions on biotite rather

* Copper hydrobiotite was synthesized during the investigation of copper bearing vermiculites from Northern Rhodesia.

than the action of potassium on vermiculite. The solutions which produced the natural hydrobiotite were probably even more dilute and required greater lengths of time than the most dilute solutions which altered biotite to vermiculite artificially.

It is possible that hydrobiotite formed as a result of slight fluctuations in the potassium to magnesium (and calcium) ratio of the altering solutions. Such fluctuations could result from fluctuations in flow and temperature of the solutions.

The following mechanism is offered as an explanation for the stability of the 1:1 regularly interstratified biotite-vermiculite:

Potassium binds the silicate layers together more tightly than hydrated magnesium or calcium because the bond length from the silicate sheet to the potassium ions in the biotite structure is shorter than the bond length from the silicate sheet to the magnesium or calcium ions in the vermiculite structure. If the potassium ions in a single layer of biotite are replaced by hydrated magnesium or calcium, then the bonding in that layer becomes weaker. As a result, the bonding in the adjacent layers which still contain potassium ions becomes stronger. The next layers of potassium ions to be replaced by hydrated magnesium or calcium are not the adjacent layers (because they are now more tightly bound) but the layers beyond the adjacent layers. When these are replaced by magnesium or calcium, the bonding is weakened and the potassium in the biotite layer, now sandwiched between two vermiculite layers, is even more tightly bound.

CONCLUSIONS

Hydrobiotite and vermiculite are believed to have formed by the supergene alteration of biotite:

1. The occurrence of domains of unaltered biotite spatially removed from domains of unaltered augite is evidence for two stages of alteration, one which produced biotite from augite and another which produced hydrobiotite and vermiculite from biotite.

2. Ion exchange experiments show that biotite readily alters to vermiculite in solutions of 0.001 molar concentration of magnesium and calcium at room temperature.

3. Ion exchange experiments show that when the potassium concentration of a solution exceeds 0.04 molar, biotite does not alter to vermiculite even when the magnesium concentration is molar and the temperature is 100° C.

4. Roy and Romo (1957) have conducted experiments which indicate that vermiculite partially alters to chlorite under hydrothermal conditions (10,000 p.s.i. and greater than 200° C.). The Libby vermiculite is not chloritic.

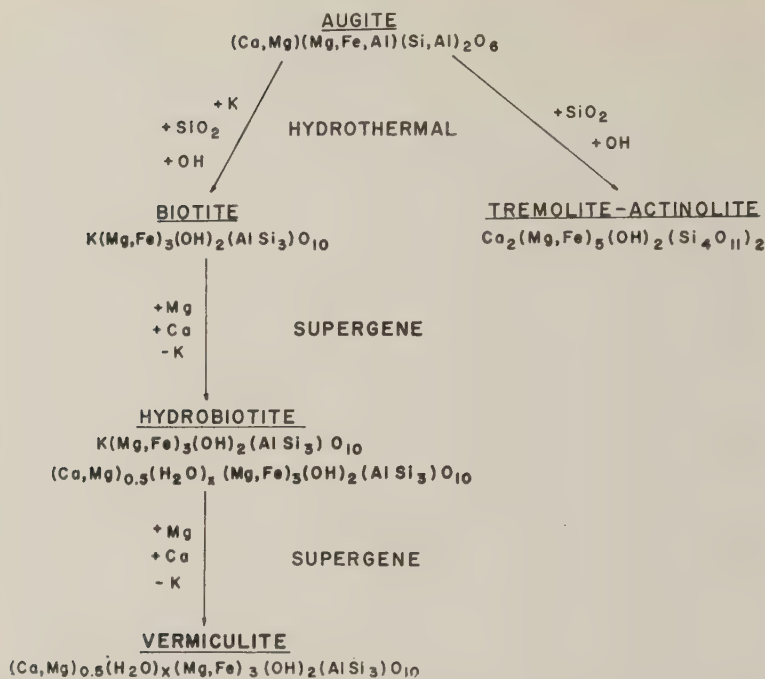


FIG. 10. The chemical changes involved in the alteration of augite at Libby, Montana.

5. Hydrobiotite, a regularly interstratified biotite-vermiculite, has not been produced from biotite in the laboratory by immersion in magnesium and calcium solutions but has been produced by immersing biotite in a molar cupric chloride solution at 100° C. for 100 hours.

The sequence of events leading to the formation of the commercial deposit of "vermiculite" at Libby, Montana started with the intrusions of a pyroxenite magma into the Belt series sediments. Textural and structural features of the pyroxenite indicate that the magma was very fluid and formed a laccolith-like body. After the pyroxenite had solidified, syenite intruded the sediments to the southwest of the pyroxenite body and penetrated the pyroxenite in the form of numerous small dikes. The syenite probably intruded a short time after the pyroxenite and was genetically related to it. At the same time that the syenite dikes intruded the pyroxenite, hydrothermal solutions rich in silica and alkalis permeated the pyroxenite, altering the augite to tremolite-actinolite and biotite.

The next important event in the formation of the deposit was probably the elevation and dissection of the region. This permitted access of supergene solutions to the altered pyroxenite even at considerable depth.

As these solutions passed through the body, they picked up magnesium and calcium from the augite or from soluble calcium and magnesium minerals which formed from augite at the time it was altered to biotite and tremolite-actinolite. The magnesium and calcium then replaced the potassium in the intersilicate layers of the biotite, altering the biotite to hydrobiotite and vermiculite. Figure 10 is a graphic representation of the chemical changes involved in this sequence of events.

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GEOCHEMICAL AND X-RAY INVESTIGATION OF NATURAL AND SYNTHETIC ANALCITES*

PRASENJIT SAHA, *The Pennsylvania State University
University Park, Pennsylvania.*

ABSTRACT

During phase-equilibria studies in the subsolidus region of the system NaAlSiO_4 (nepheline)- $\text{NaAlSi}_3\text{O}_8$ (albite)- H_2O , analcite was synthesized from glasses of a wide range of composition. Geochemical, optical, and x-ray investigations were carried out and it was found that there is a wide range of solid solution of analcite. Determination of water content and unit cell constant proved that the variation of these properties are linear functions of the silica content of the analcites. The synthetic analcites were found to be isotropic, but the refractive index variation was not found to be linear. The experimental data have been discussed in the light of previous structural investigations of analcites. Data on the composition of natural analcites have been collected from literature and summarized.

INTRODUCTION

During a study of the phase equilibrium relations in the system $\text{NaAlSi}_3\text{O}_8$ - NaAlSiO_4 - H_2O analcite was synthesized from glasses ranging in composition from NaAlSiO_4 to $\text{NaAlSi}_3\text{O}_8$. The present paper deals with some optical and powder diffraction studies of the synthetic and some natural analcites. Data from the literature have also been collected and summarized.

The structure of analcite ($\text{NaAlSi}_2\text{O}_6 \cdot \text{H}_2\text{O}$) has been worked out by W. H. Taylor (1930, 1938). It is a zeolite, the main structural feature being the presence of diagonal channels surrounded by six-fold rings of silica tetrahedra; the channels do not cross each other. The water molecules are situated inside these channels. The sodium atoms are surrounded by four-fold rings of silica tetrahedra. The structure is apparently cubic, and the lattice constant has been determined to be 13.7 \AA (Bragg, 1937), but some variation has been reported (Grüner, 1928). The space group has been determined to be $Ia3d(O_h^{10})$, but Grüner suggested that $Im3m(O_h^9)$ is more probable, since he observed some reflections in oscillation photographs which could not be accounted for by the space group $Ia3d$. Schiebold (1930) considered the mineral to be tetragonal, space group $I4/acd(D_{4h}^{20})$. Taylor assumed equivalent positions for Si and Al, assigned the water molecules to a 16-fold special position (0.125, 0.125, 0.125), but assumed that the 16 Na atoms randomly occupied a 24-fold special position (0.125, 0, 0.25). Náray-Szabó (1938) suggested that analcite has the same structure as pollucite, i.e., space group D_{4h}^{20} pseudomorphic after O_h^{10} , and he proposed an alternate arrangement for the Na atoms and the water molecules, the former oc-

* Mineral Industries Contribution No. 58-3.

cupying one 16-fold special position (0.125, 0.125, 0.125) and the latter partially occupying a 24-fold special position (0, 0.25, 0.125).

All these structural investigations were based on the following assumptions: (1) Ideal, stoichiometric composition of analcite was assumed ($\text{NaAlSi}_2\text{O}_6 \cdot \text{H}_2\text{O}$), (2) though x-ray photographs of analcite indicate cubic symmetry, many of the specimens examined show optical anomalies, i.e., weak to moderate birefringence. Hence it has been assumed that the mineral is pseudo-cubic. As Bragg (1937) states, "It is not definitely established that statistically cubic symmetry is brought about by the constant interchange of the Na atoms. Whereas perfect cubic symmetry is impossible with 16 fixed Na atoms distributed among 24-fold positions, Taylor points out that such a distribution is compatible with tetragonal symmetry. It is possible that analcite has an intimate twinning of tetragonal components which simulates cubic symmetry."

Barrer (1950) has shown that the sodium ions in analcite can be readily exchanged by potassium, thallium, and rubidium ions. If Taylor's model of analcite structure is assumed to be correct, it is difficult to explain how these ions can substitute for sodium. The Na-O distance (Beattie, 1954), in the analcite structure, assuming Taylor's model, is 2.6 Å, leaving only 1.2 Å for the cationic radius, if the oxygen is taken to have an ionic radius of 1.4 Å. Thus potassium, thallium, and rubidium ions are too large to fit into the sodium positions. All these ion-exchanged analcites are noncubic at room temperature. Beattie (1954) explained this kind of substitution by assuming that potassium, thallium, and rubidium ions will be repulsed and forced away from the sodium position towards the position normally occupied by caesium in pollucite (0.125, 0.125, 0.125). The potential energy for such an ion will then lie in some position intermediate between that of caesium and that of sodium. The structure will then no longer possess cubic symmetry.

Recently Coombs (1955) obtained some powder data on noncubic analcite. The powder photograph shows distinct evidence of the noncubic nature of the specimen. Coombs has classified analcite into the three following groups: (1) Strictly cubic and isotropic, space group $Ia\bar{3}d$. The synthetic specimen, hydrothermally prepared (Barrer, 1952), appears to belong here, (2) birefringent and biaxial, departures from cubic lattice dimensions not detectable although anomalous x-ray reflections may appear, i.e., Flinders analcime, (3) trigonal or nearly trigonal with rhombohedral angle approximately 90.5° , e.g., Låvan analcime. Biaxial specimens from the same locality must be of lower symmetry.

Single crystal work has not been done with the distinctly noncubic Låven specimen.

Guyer, Ineichen, and Guyer (1957) reported the synthesis of analcites

from compositions ranging from $\text{Na}_2\text{Al}_2\text{Si}_3\text{O}_{10}$ (anhydrous natrolite) to $\text{NaAlSi}_5\text{O}_{12}$ (anhydrous sodium mordenite). The powder diffraction data presented in the paper show distinct evidence of change in the lattice constant. The unit cell size decreases towards higher silica compositions. They assumed that this is due to the smaller bond distance between silicon and oxygen atoms compared to the Al-O bond distance.

PRESENT WORK

Subsolidus studies in the system $\text{NaAlSi}_3\text{O}_8$ - NaAlSiO_4 - H_2O indicate that analcite crystallizes from glasses of composition ranging from NaAlSiO_4 (nepheline) to $\text{NaAlSi}_3\text{O}_8$ (albite). The phase equilibria studies will be discussed in a later paper. The most interesting feature is that analcite is the only phase which crystallizes from glasses of composition ranging from $\text{Na}_2\text{Al}_2\text{Si}_3\text{O}_{10}$ (anhydrous natrolite) to $\text{NaAlSi}_3\text{O}_8$ (albite), except for very small quantities of another yet unidentified phase in some runs ($\sim 3\%$, modal analysis). When observed under the microscope, no unreacted glass could be observed in these runs. Since all the runs were made in sealed gold tubes, the analcite crystals synthesized must have the same composition as the glass from which it crystallized, with the addition of water.

Optical Properties of Analcite Solid Solution

The synthetic analcites were found to be isotropic, quite in accord with Barrer's (1952) and Coombs' (1955) observations. The small amount of the other still unidentified phase in some runs has a high refringence and birefringence, and can be readily distinguished from the analcite grains. The refractive index of synthetic analcites of different compositions have been plotted in Fig. 1 and show that the refractive indices decrease with increasing silica content. The limit of error is $\pm .002$.

X-Ray Diffraction Studies of Analcite Solid Solution

The variation in composition of synthetic analcites was confirmed by the change in d -spacings observed in the back reflection region, using silicon as internal standard. The (639) peak of analcite falls at higher angles for analcites of higher silica content. The resolution of the $K\alpha_1$ and $K\alpha_2$ peaks in analcite of albite composition is not so well defined, and this is probably due to imperfect crystallization. The difference $2\theta_{\text{An}(639)} - 2\theta_{\text{Si}(331)}$ ($\text{CuK}\alpha_1$ radiation) has been plotted against variation in SiO_2 content in Fig. 2. The lattice constant " a " was calculated from the change in d -spacing and plotted in Fig. 3. Both Figs. 2 and 3 show that the variations are linear.

Hydrothermal studies on the stability of natrolite showed that only

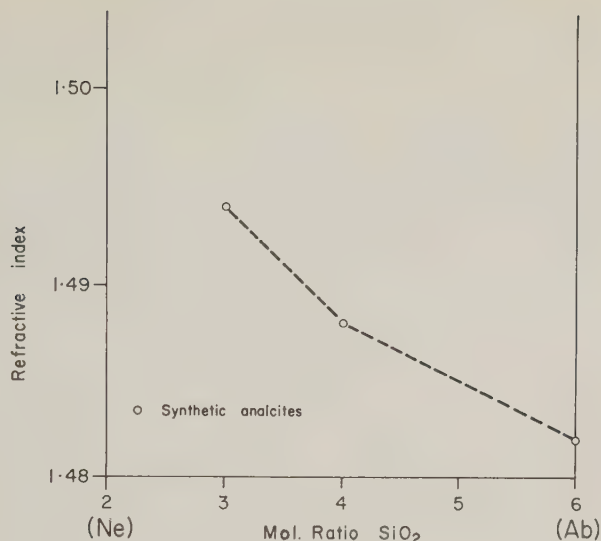


FIG. 1. Refractive index of synthetic analcites.

analcite appeared in glasses of anhydrous natrolite composition ($\text{Na}_2\text{Al}_2\text{Si}_3\text{O}_{10}$). Runs made with natural specimens of natrolite also indicated that natrolite breaks down to analcite. The d -spacing variation of this analcite has been plotted in Fig. 2, and assuming stoichiometric composition of the natrolite, it can be seen that it falls very close to the curve.

Two specimens of analcite from sedimentary rocks (Yavapai County, Arizona) were kindly sent by Dr. C. S. Ross for x -ray diffraction and single crystal studies. These analcites have previously been described by Ross (1928). The crystals are isotropic and the refractive index is 1.483. The $K\alpha_1$ peak is sharp, but the $K\alpha_2$ peak is not so well resolved. However single crystal work shows that there is no deviation from cubic symmetry. In his paper (1928) Dr. Ross gave two analyses. One interesting feature of these analyses is that the water and silica content are in excess of that required for the ideal composition of analcite. There is no appreciable difference in the proportion of silica in the two analyses. The author received two samples of these analcites, but did not know which specimen represented which analysis. The d -spacing variations were plotted assuming the mean of the two analyses. The points fall very close to the curve for the synthetic analcites (Fig. 2). The small differences in the silica contents of the two analyses will not significantly change the position of the points on this graph.

A specimen of analcite occurring as clusters of transparent icositetra-

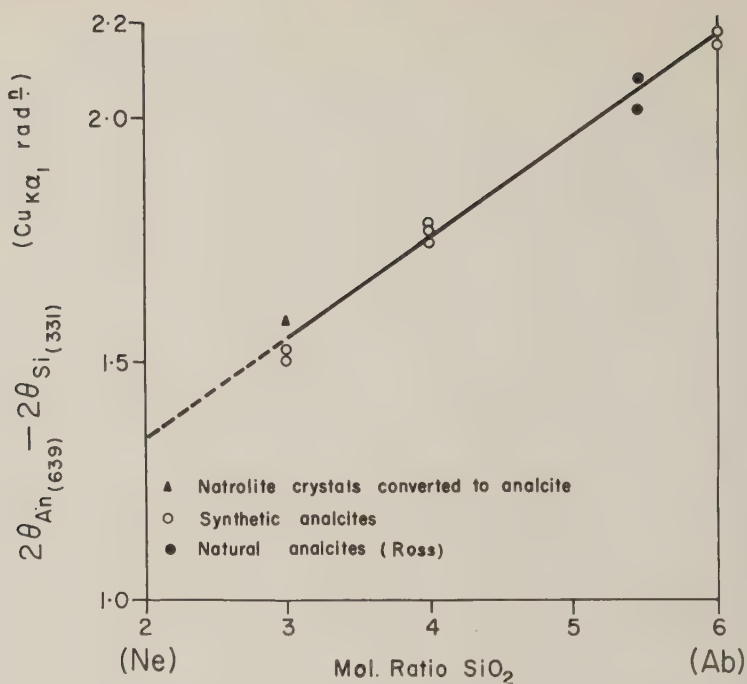


FIG. 2. Variation of $d_{(639)}$ spacing of synthetic and natural analcites with variation of SiO_2 content.

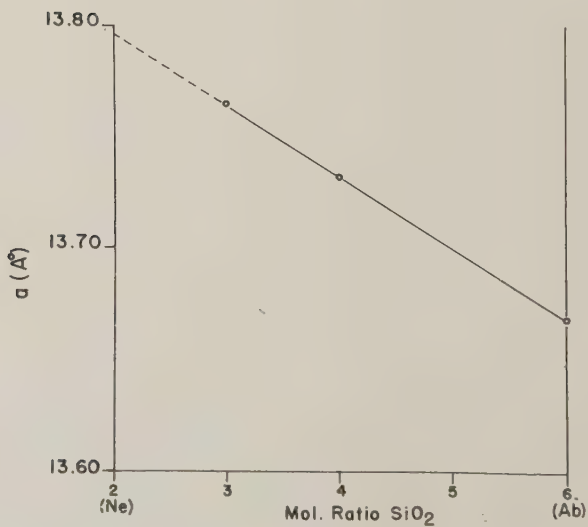


FIG. 3. Variation, with composition, of the lattice parameter of synthetic analcites.

hedral crystals in the cavities of a basaltic rock (Specimen No. 450.2, Patagonia, Sicily) indicates a relatively small shift in the (639) peak compared to the Si internal standard. The mineral is weakly anisotropic, and the refractive index is $1.488 \pm .002$. Single crystal photographs indicate that the specimen is cubic. Though the sample has not yet been analyzed, a deficiency in water and silica is suspected. The main purpose of including this sample in this discussion was to bring out the shifting of the (639) peak for different natural analcites.

The x-ray diffraction pattern of a specimen of analcite from Patterson, New Jersey, which was found to be moderately birefringent gave a very diffuse (639) peak in the back-reflection regions.

Geochemical Studies of Natural Analcites

Data on the variation of chemical composition of analcite were collected from the literature. Sixty-eight analyses were collected from Doelter (1917), and 25 were collected from other papers, summarized in the Mineralogical Abstracts. In the zeolite structure, as in other tectosilicate structures, the charge deficiency caused by the substitution of aluminum ions for silicon ions is compensated by larger cations like calcium, sodium and potassium.* Thus for charge balance, $\text{Al} = 2\text{Ca} + \text{Na} + \text{K}$. This assumption was used as a check on the accuracy of the analyses. A deviation of ± 30 atoms was allowed. All the available analyses from the literature have been tabulated in Table 1, and the mol. ratios of silica and water have been calculated, taking the arithmetic mean of the number of Al atoms and $(2\text{Ca} + \text{Na} + \text{K})$ atoms as unity. On this basis, 59 analyses could be plotted (Fig. 4). One very significant feature is apparent from the diagram, the water content increases with the increase in silica content.

Analyses of Water of Synthetic Analcites

The water content of analcites synthesized from glasses of compositions $\text{Na}_2\text{Al}_2\text{Si}_3\text{O}_{10}$, $\text{NaAlSi}_2\text{O}_6$, and $\text{NaAlSi}_3\text{O}_8$ were determined by the loss on ignition method. About 50 milligrams of analcites of these compositions were prepared hydrothermally in runs of 1 month duration. The charges were sealed in thick-walled gold tubes with excess water, so that nothing could be carried away in solution. The gold tubes were carefully weighed before and after the synthesis, and since there was no appreciable loss in weight the composition of the charges did not change. After the synthesis, a small portion of the product was observed under the

* Since calcium is divalent, two aluminum ions are required to substitute for two silicon ions when one calcium ion enters the structure. Substitution of one aluminum ion for one silicon ion is required when one sodium or potassium ion enters the structure.

TABLE I. ANALYSES OF ANALCITE

No.	Number of atoms							SiO ₂ :H ₂ OX Mol. ratio
	Si	Al	H ₂ O	Ca	Na	K	2Ca+Na +K Δ ⁻	
1	918	451	459	—	437	—	437 14	4.1:2.1
2	940	431	489	—	445	—	445 14	4.3:2.2
3	917	436	456	—	442	—	442 6	4.2:2.1
4	854	462	586	—	166	—	166 296*	
5	926	451	438	—	473	—	473 22	4.0:1.9
6	955	443	500	6	383	12	407 36*	
7	918	462	458	—	459	—	459 3	4.0:2.0
8	957	454	444	100	208	—	408 46*	
9	934	471	444	104	208	—	416 55*	
10	915	463	460	—	454	—	454 9	4.0:2.0
11	917	454	453	—	454	—	454 0	4.0:2.0
12	894	452	472	22	256	95	395 57*	
13	919	454	426	5	393	32	435 19	4.1:1.9
14	905	463	450	4	420	14	442 21	4.0:2.0
15	936	436	462	5	390	31	431 5	4.3:2.1
16	944	416	544	—	294	—	294 122*	
17	927	441	488	—	398	—	398 43*	
18	912	453	461	4	419	17	444 9	4.1:2.1
19	913	443	503	16	406	—	438 5	4.2:2.3
20	907	451	460	—	452	—	452 1	4.0:2.0
21	925	437	475	—	444	—	444 7	4.2:2.2
22	929	481	488	—	413	—	413 68*	
23	889	455	501	—	468	3	471 16	3.8:2.2
24	938	432	488	9	426	—	444 12	4.3:2.2
25	893	451	474	7	433	—	447 4	4.0:2.1
26	887	458	470	11	469	—	491 33*	
27 ¹	898	483	472	—	717	28	745 262*	
28	912	464	473	6	436	—	448 16	4.0:2.1
29	906	449	454	30	341	43	444 5	4.1:2.0
30	899	471	465	—	436	—	436 35*	
31a	906	449	451	16	419	—	451 2	4.0:2.0
b	920	416	278	66	288	58	478 62*	
32	950	422	462	—	440	7	447 25	4.4:2.1
33	912	449	462	—	452	—	452 3	4.1:2.1
34	911	454	458	6	397	17	426 28	4.1:2.1
35	929	440	465	—	435	—	435 5	4.2:2.1
36	907	457	453	—	454	—	454 3	4.0:2.0
37	892	472	460	15	439	—	469 3	3.8:2.0
38	894	473	463	8	486	—	502 29	3.7:1.9
39	883	463	444	—	491	—	491 28	3.7:1.9
40	886	486	458	—	472	—	472 14	3.7:1.9
41a	914	457	416	34	336	34	438 19	4.1:1.9
b	830	442	613	47	352	57	503 61*	
42	896	461	450	—	466	—	465 5	3.9:1.9
43a	954	410	509	6	369	—	381 29	4.8:2.6
b	968	401	488	7	369	—	383 18	4.9:2.5
c	973	423	455	6	358	—	370 53*	
d	981	391	492	6	376	—	388 3	5.0:2.5
e	934	425	500	14	369	—	397 28	4.5:2.4
44	919	448	477	—	434	—	434 14	4.2:2.2
45	906	452	470	4	430	4	442 10	4.1:2.1
46	853	471	504	29	375	27	460 11	3.7:2.2
47 ²	950	421	465	3	394	—	400 21	4.6:2.3
48 ²	928	452	458	3	402	—	408 41*	

TABLE 1 (Continued)

No.	Number of atoms							Δ^-	$\text{SiO}_2:\text{H}_2\text{O}\times$ Mol. ratio
	Si	Al	H ₂ O	Ca	Na	K	2Ca+Na +K		
49	943	402	516	—	407	—	407	5	4.7:2.6
50	879	424	455	12	477 ^a	—	501	77*	
51	702	604	696	148	197	—	493	111*	
52	913	475	472	1	387	—	389	87*	
53	906	426	483	24	444	—	492	66*	4.3:2.1
54	830	201	791	223	356	21	823	622*	
55	941	430	485	10	367	11	398	32*	
56	936	425	466	6	429	9	450	25	
57	941	438	481	—	421	—	421	17	4.4:2.2
58	931	440	463	—	417	—	417	23	4.4:2.2
59	906	463	457	—	431	3	434	29	4.1:2.0
60	910	458	470	—	435	—	435	23	4.1:2.1
61	917	448	465	—	425	—	425	23	4.2:2.1
62	940	434	466	—	421	2	423	11	4.4:2.2
63	946	448	459	—	409	5	414	34*	
64	894	450	507	—	460 ⁴	—	460	10	3.9:2.0
65	902	464	457	8	432	7	455	9	
66	899	456	456	31	356	34	452	4	
67	902	454	455	28	357	34	447	7	
68 ²	899	484	479	17	324	48	406	78*	4.0:2.0
69	904	464	464	—	445	—	445	19	
70	978	416	466	2	368	2	376	40*	
71	942	438	483	3	385	9	400	38*	
72 ³	856	452	533	21	274	95	411	41*	4.1:2.0
73	914	457	417 ⁵	34	337	34	439	18	
74	830	442	614 ⁵	47	352	57	502	60*	
75	941	424	452 ⁶	14	322	47	397	27*	
76	918	449	456	2	445	4	453	4	4.0:2.0
77	926	498	472	2 ⁷	352	3	359	139*	
78	907	463	456	—	435	4	439	24	
79	1,009	353	464	1	354	21	377	24	
80	979	351	477	9	316	32	366	15	5.4:2.7
81	957	409	468	9	353	14	385	24	4.8:2.4
82	909	452	483	8	436	—	452	0	4.0:2.1
83	905	428	501	21	377	23	442	14	4.2:2.3
84	881	483	425	3	429	16	451	32*	4.4:2.3
85	940	436	482	—	419	—	419	17	
86 ^a	1,000	388	464	4	374	2	384	4 ₈	
<i>b</i>	1,094	334	370	14	305	2	335	1 ₈	

$\Delta^- = |\text{Al} - (2\text{Ca} + \text{Na} + \text{K})|$.

^x—Taking the arithmetic average of the number of Al atoms and the number of (2Ca+Na+K) atoms as unity.

*—Analyses not used for plotting in Fig. 4.

1—Addition of oxides does not give the total stated in the analysis.

2—Only (H₂O+) plotted in the column marked "H₂O." In the other analyses, where only "H₂O" is stated in the analyses, it was assumed that all the water is (H₂O+) water.

3—Na+K.

4—Not plotted in Fig. 4, because wt% Na₂O was assumed to be = 100 - Σ other constituents (SiO₂+Al₂O₃+H₂O).

5—(H₂O+)+(H₂O-). Not plotted in Fig. 4.

6—(H₂O+) and (H₂O-) probably interchanged by mistake in the original analysis. Not plotted in Fig. 4.

7—(CaO+SrO).

8—Not plotted because of enclosed grains of quartz.

TABLE 1 (*Continued*)

No.	Year	Name of analyst	Locality	Association
1	1822	Rose	Fassatal, Tyrol	—
2	1822	Rose	Fassatal, Tyrol	—
3	1829	Connel	Old Kilpatrick, Scotland	—
4	1836	Thomson	Dumbarton, Kilpatrick, Scotland	(weathered, cloudy sp.)
5	1836	Thomson	Giant's Causeway, Ireland	Amygdaloidal basalt
6	1839	Henry	Blagodat Mt., Ural	(with magnetite)
7	1842	Awdejew	Lövö, Norway	—
8	1847	Riegel	Niederkirchen, Bavaria	—
9	1847	Riegel	Niederkirchen, Bavaria	—
10	1850	Bork & Berlin	Låven, Norway	—
11	1850	Bork & Berlin	Låven, Norway	—
12	1853	Sartorius v. Walters- hausen	Cyclopean Islands, Sicily	Volcanic rock's
13	1858	Rammelsberg	Cyclopean Islands, Sicily	—
14	1858	Rammelsberg	Cyclopean Islands, Sicily	—
15	1858	Rammelsberg	Wesseln, Bohemia	—
16	1863	Stromeyer	Duingen, Hannover	Ironstone shale (?)
17	1864	Bischof	Seiseralpe, Tyrol	—
18	1869	Tschermak	Punzau, Silesia	Teschenite
19	1873	Young	Crofthead, Scotland	—
20	1873	Young	Mugdock, Scotland	—
21	1873	Young	Barrhead, Scotland	—
22	1874	Pisani	Brevik, Norway	—
23	1875	Paijkull	Brevik, Norway	—
24	1876	Lemberg	Fassatal, Tyrol	Melaphyre
25	1877	Lemberg	Predazzo, Tyrol	Porphyritic rock
26	1878	Harrington	Montreal, Canada	—
27	1879	Luedecke	Heldburg, Thuringia	In fissures in phonolith
28	1880	Preis	Kuchelbad, Bohemia	In druses in diabase
29	1881	Ricciardi & Speciale	Cyclopean Islands, Sicily	Basalt
30	1881	Damour	Låven, Norway	—
31a	1881	Nikolajew	Blagodat, Ural	(crystalline sp.)
b	1881	Nikolajew	Blagodat, Ural	(massive sp.)
32	1882	Bamberger	Monte Catini, Tuscany	With chalcopyrite in gabbro
33	1883	Lemberg	Låven, Norway	—
34	1884	Sauer	Wiesental, Bohemia	Leucitophyre
35	1885	Cross & Hillebrand	Table Mt., Colorado, U.S.A.	In vesicles in basalt
36	1884	Lorenzen	Kangerdluarsuk, Greenland	—
37	1887	Hersch	Cyclopean Islands, Sicily	—
38	1889	Johnsson	Klein-Arö, Norway	—
39	1889	Langlet	Klein-Arö, Norway	—
40	1889	Brögger	Eikaholmen, Norway	—
41a	1890	Lindgren	Highwoods, Rocky Mt., Montana, U.S.A.	Sandstone
b	1890	Melville	Highwoods, Rocky Mt., Montana, U.S.A.	Sandstone
42	1892	Brauns	Friedensdorf, Hessen	In fissures in diabase
43a	1893	Zschau	Dresden, Saxony	Syenite (colorless sp.)
b	1893	Zschau	Dresden, Saxony	Syenite (red sp.)
c	1893	Zschau	Dresden, Saxony	Syenite (red sp.)
d	1893	Zschau	Dresden, Saxony	Syenite (red sp.)
e	1893	Zschau	Dresden, Saxony	Syenite (colorless sp.)
44	1895	Glinka	Kobi, Persia	Mud volcano
45	1896	Fairbanks	Point Sal, California, U.S.A.	Augite teschenite
46	1895	Hillebrand	Colorado, U.S.A.	Basalt
47	1899	Clarke & Steiger	Wasson's Bluff, Nova Scotia	—
48	1900	Clarke & Steiger	North Table Mt., Colorado, U.S.A.	—
49	1901	Thugutt	Seiseralpe, Tyrol	—
50	1901	Evans	Mt. Girnar, Kathiwar, India	Monchiquite

TABLE 1 (Continued)

No.	Year	Name of analyst	Locality	Association
51	1904	Takimoto	Maze, Echigo, Japan	—
52	1905	Egleson	Montreal, Canada	Nepheline syenite
53	1907	Anderson	Ben Lomond, Australia	—
54	1907	Tsukamoto	Maze, Echigo, Japan	Basalt
55	1908	Baschieri	Seiseralpe, Tyrol	—
56	1907	Manasse	Hamasat, Massaua	Augite teschenite
57	1913	Tschirwinsky & Orloff	Kara-Dagh, Krim	—
58	1912	Foote & Bradley	Two Isles, N.S.	(average of two sp.)
59	1912	Foote & Bradley	Cyclopean Islands, Sicily	(average of two sp.)
60	1912	Foote & Bradley	Kerguelen Islands	(average of two sp.)
61	1912	Foote & Bradley	Victoria, Australia	(average of two sp.)
62	1912	Foote & Bradley	Michigan, U.S.A.	(average of two sp.)
63	1912	Foote & Bradley	Montreal, Canada	(average of two sp.)
64	1928	Curlo	Murlo, Tuscany	—
65	1926	Di Franco	Cyclopean Islands, Sicily	Basalt
66	1929	Grassi-Cristaldi & Scafile	Cyclopean Islands, Sicily	—
67	1929	Grassi-Cristaldi & Scafile	Cyclopean Islands, Sicily	—
68	1928	Hewett, Shannon & Gonyer	Ritter Hot Spring, Grant County, Oregon, U.S.A.	In cavities in vesicular basalt
69	1929	Hodge-Smith	Kyogle, Australia	In vesicles in basalt
70	1939	Kašper	Morcinov, Czechoslovakia (?)	Amygdaloidal melaphyre
71	1933	Kratochvíl	Budnany, Bohemia	Diabase
72	1938	Larsen & Buie	Highwood Mt., U.S.A.	Phenocrysts in analcite basalt
73	1938	Larsen & Buie	Highwood Mt., U.S.A.	Dyke rock
74	1938	Larsen & Buie	Highwood Mt., U.S.A.	Dyke rock
75	1928	Lonsdale	Terlingua, U.S.A.	Altered basalt
76	1945	Morgante	Diredana, Eritrea	Basalt lavas & tuffs
77	1935	Reichert & Erdélyi	Dunabogdany, Ungarn	—
78	1947	Roques	Kassa Island	Nepheline syenite pegmatite
79	1928	Ross	Yavapai, Ariz., U.S.A.	Lake beds & playa deposits
80	1928	Ross	Yavapai, Ariz., U.S.A.	Lake beds & playa deposits
81	1924	Ross & Shannon	Challis, Idaho, U.S.A.	Lake beds & andesite flows
82	1915	Shimizu	Maze, Echigo, Japan	—
83	1924	Smirnov	Mt. Imeretin, Trans-Caucasia	Andesite
84	1941	Stewart	Loch Borolan, Scotland	Pegmatitic patch in borolanite
85	1935	Tiselius	Faeroe Island	—
85a	1926	Zeberg	Chaitsyn Cape, Arctic Russia	Augite porphyrite (clear outer portion—av. of 5 sp.)
85b	1926	Zeberg	Chaitsyn Cape, Arctic Russia	Augite porphyrite (cloudy inner portion—av. of 4 sp.) Inclusions of quartz present

microscope. Only analcite and a small amount of the unidentified phase were found, but no glass. The rest of the samples were taken out of the tubes, and then carefully dried at room temperature in a well-sealed desiccator containing phosphorus pentoxide as the drying agent, special care being taken to change the phosphorus pentoxide whenever a glassy coating of metaphosphoric acid was formed (Hillebrand, 1953). The process of weighing was continued for about a week, until the weights became constant. The samples were then heated to 900° C. and weighed.

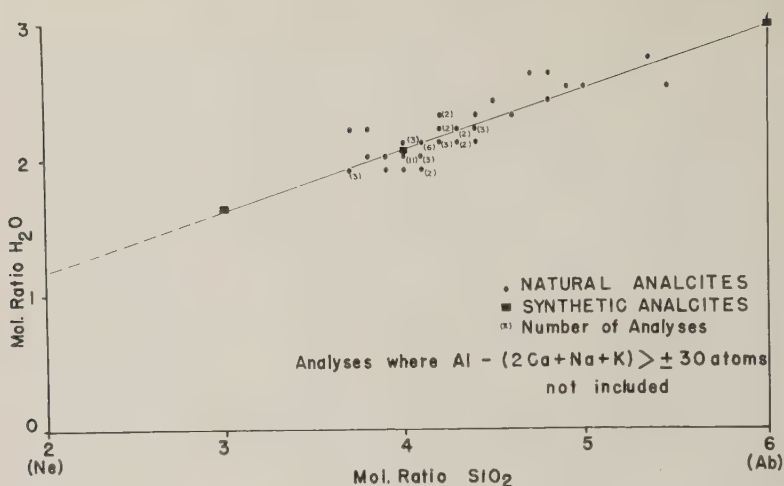


FIG. 4. The water content of synthetic and natural analcites plotted against silica.

The difference in weight was assumed to be the weight of structural water. The analyses, neglecting the small amount of the unidentified phase, are tabulated below:

Composition of glass:	$\text{Na}_2\text{Al}_2\text{Si}_3\text{O}_{10}$	$\text{NaAlSi}_2\text{O}_6$	$\text{NaAlSi}_3\text{O}_8$
Mol. ratio water:	1.6 ± 0.1	2.0 ± 0.1	2.9 ± 0.1

The analyses, when plotted in Fig. 4, indicated that the water content of analcrite varies linearly with the variation in silica. This is in accord with the composition of natural analcites, as indicated in the figure, except the range of composition of natural analcites is more restricted.

CONCLUSION

From the present work on the composition of analcites, it becomes apparent that the analcrite solid solution can be represented as follows:

$\text{NaAlSi}_{1.5}\text{O}_5, 0.75\text{H}_2\text{O}$	$\text{NaAlSi}_2\text{O}_6, \text{H}_2\text{O}$	$\text{NaAlSi}_3\text{O}_8, 1.5\text{H}_2\text{O}$
Na:Al:Si:H ₂ O	Na:Al:Si:H ₂ O	Na:Al:Si:H ₂ O
1:1:1.5:0.75	1:1:2:1	1:1:3:1.5
(natrolite composition)	(ideal analcrite composition)	(albite composition)

Taylor's model indicates $48[(\text{Si}, \text{Al})\text{O}_2]$ in the unit cell. Thus there are 96 oxygen atoms and 48 Al+Si atoms in the unit cell. This number of Al+Si atoms, and consequently the number of oxygen atoms, must remain constant in order to avoid any gross distortion of the lattice—which should manifest itself in a change in space group but which has not been observed in single crystal or powder diffraction work—since the whole

framework of the lattice is built up by connecting silica and alumina tetrahedra. Thus by keeping the number of oxygen atoms constant,* we can calculate the number of other atoms and molecules in the unit cell of analcite in the following way:

Composition	O	Al	Si	Al+Si	Na	H ₂ O
NaAlSi ₄ O ₆ 0.75H ₂ O	96	19.2	28.8	48	19.2	14.4
NaAlSi ₂ O ₆ , H ₂ O	96	16	32	48	16	16
NaAlSi ₃ O ₈ , 1.5H ₂ O	96	12	36	48	12	18

* Assuming (OH)⁻ not substituting for O²⁻.

Assuming Taylor's model of analcite structure to be correct, the distribution of the 19.2 sodium atoms and 14.4 water molecules found in natrolite composition analcite is reasonable since a 24-fold position for the sodium atoms and a 16-fold position for the water molecules are available. However, it is impossible to distribute 18 H₂O molecules in albite composition analcite in a 16-fold position. An alternative suggestion, that the excess water molecules lie in some position intermediate between that of caesium in pollucite and sodium in analcite seems to be unlikely, since the albite composition analcites are fully hydrated, isotropic under the microscope, and x-ray diffraction patterns reveal a cubic symmetry. Another alternative is to assume that the excess water, i.e., the water in excess of 16 molecules, occupies some vacant sodium positions. A consideration of the radius of the water molecule, as in the case of rubidium and thallium ions (Beattie, 1954) makes it somewhat improbable, but in the absence of an alternative suggestion, this seems to be the most likely possibility since the water molecule is not as large as the rubidium or thallium ion, and only two excess water molecules have to be accommodated in the structure. It will be, perhaps, almost impossible to detect such slight distortion of the structure by x-ray methods.

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GRAPHS FOR THE ELIMINATION OF THE HARTMANN NET IN THE DETERMINATION OF REFRACTIVE INDICES IN HIGH DISPERSION MEDIA

JOEL S. WATKINS, JR., *University of Wisconsin, Madison, Wisconsin**

ABSTRACT

Direct reading charts are presented which eliminate graphical solutions (Hartmann Nets) required for index of refraction determinations in high dispersion media. Geometric proofs and construction procedures are explained. A special chart adapted to high dispersion liquids based on Tsuboi's classic work determines the composition of plagioclase feldspars from a single determination of N_x' .

INTRODUCTION

The determination of indices of refraction in high dispersion liquids requires a certain amount of graphing to obtain a solution to the problem. A single chart can be made which gives directly a simple solution to the problem.

A discussion of the entire method of determination of indices using high dispersion liquids is beyond the scope of this paper and the reader who is unfamiliar with the techniques is referred to a discussion by Emmons and others (1928, 1948, 1953) which covers the subject in detail.

Essentially, a properly oriented mineral is immersed in a liquid having an index for sodium light near that of the mineral in question. Since the change of index with respect to the change in wavelength is much larger in the liquid than in the mineral, it is possible to find some wavelength of light where the indices of mineral and liquid are the same. This will be subsequently referred to as the "wave length of agreement," (λ^*).

If this process is repeated with a second liquid, thus finding another wave length of agreement, the index of the mineral has been established for two different wave lengths. If the indices of the liquids are plotted on a Hartmann Net as a function of wave length, the wave lengths of agreement can be used to determine the sodium light index as shown diagrammatically in Fig. 1.

The values $N_D(\text{mineral})-N_D$ (liquid No. 1) have been derived directly by computation, and are expressed in a chart for all possible values of the two wave lengths of agreement (Fig. 2). Two considerations must be satisfied before the chart is valid. (1) The lines representing the dispersion of liquids on the Hartmann Net must be essentially parallel. (2) The value read is a function of the difference in index for sodium light. Although the chart can be constructed to yield valid answers for any difference in index, this difference must be the same for all adjacent

* Present address: University of Texas, Austin, Texas.

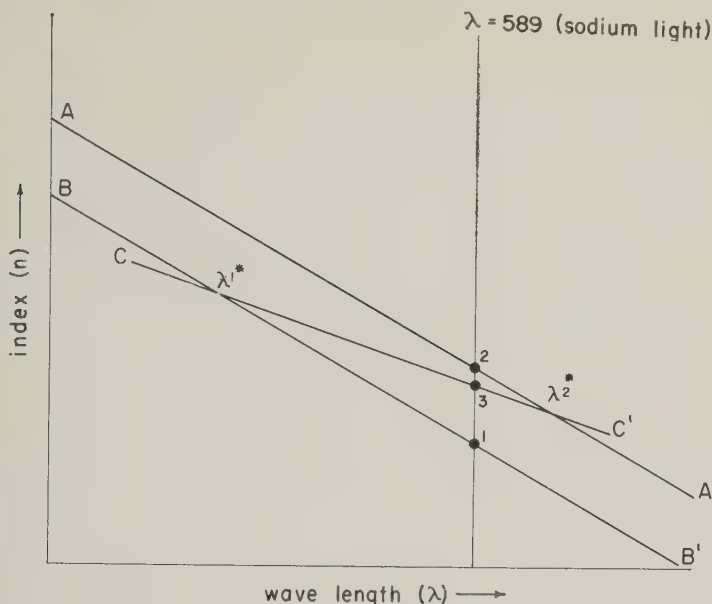


FIG. 1. Graphical solution for N_D of a mineral, using high dispersion liquids. AA' = plot of upper liquid dispersion; BB' lower liquid. λ_1^* and λ_2^* are wave lengths of agreement (1) and (2) respectively, where (1), (2) and (3) are N_D of lower liquid, upper liquid and mineral, respectively. N_D of mineral is determined by drawing a straight line from λ_1^* to λ_2^* .

liquids or the chart loses much of its usefulness. Standard immersion media are uniformly spaced, usually with intervals of 0.005, e.g., 1.500, 1.505,

USE OF THE DIAGRAM

No changes in microscope procedure are necessary for the use of the diagram. Two wave lengths of agreement are found using two adjacent liquids. The index for sodium light is known for both liquids. If the wave length of agreement for the liquid with the lower index of refraction is $510 \text{ m}\mu$. and that for the upper liquid is $610 \text{ m}\mu$., one reads up from $510 \text{ m}\mu$. on the abscissa and across from $610 \text{ m}\mu$. on the ordinate to the intersection of these two lines (Y, Fig. 2). This intersection falls between the correction lines 0.004 and 0.005 and interpolation gives in this case about 0.0042. Adding this last value to the sodium line index of the lower liquid, one obtains the correct sodium line index of the mineral.

For example, if the lower liquid has an index of 1.500 and the wave lengths of agreement are as above, then the index of the mineral for sodium light is 1.5042. If another mineral is tested and it is found to have wave lengths of agreement of $490 \text{ m}\mu$. and $570 \text{ m}\mu$. respectively (X,

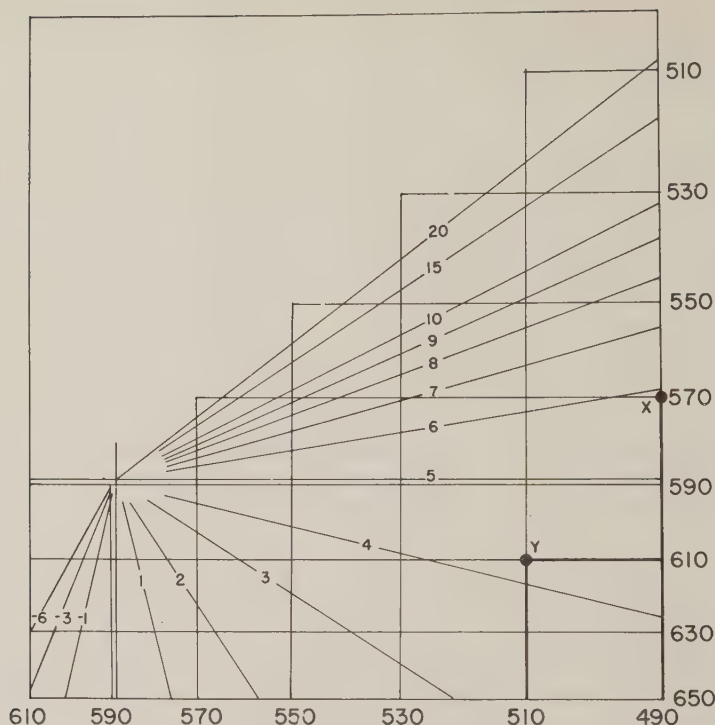


Fig. 2. N_D correction chart. Points X and Y refer to values used in examples. Correction $\times 10^3$. Thus number 4 represents a correction of 0.004.

Fig. 2), the liquids remaining the same, the correction is approximately 0.006 and the mineral has an index of 1.506 for sodium light. Figure 3 shows the same results as obtained on a Hartmann Net. It must be remembered that the examples are based on differences in index of 0.005 for the liquids. A second chart is shown in Fig. 4. This chart is read the same way as the first, except that the correction when subtracted from the mean dispersion of the two liquids gives the dispersion of the mineral, i.e., $N_F - N_C$. If the dispersion slopes of the two liquids are very nearly the same, either can be used with little error. The same conditions apply to this chart as applied to the first. (Fig. 4 as shown is not as accurate as Fig. 2; see construction below.) For example, if the dispersion of the lower liquid is 0.020 and that of the upper liquid is very nearly the same, and if the wave lengths of agreement are 510 $m\mu$. and 610 $m\mu$., then the correction is about 0.0084 and the dispersion of the mineral is equal to $(N_F - N_C) \text{ liquid} - (\text{Correction}) = 0.020 - 0.0084 = 0.0116$. This relation can be verified in Fig. 3.

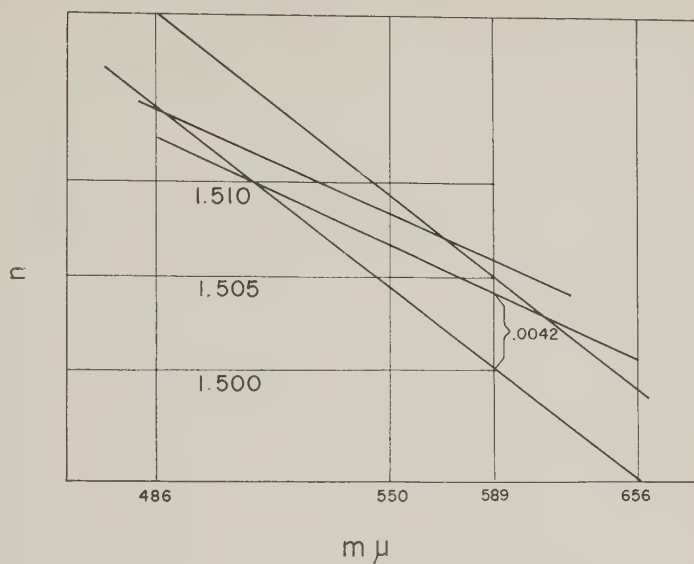


FIG. 3. Graphical solution of examples.

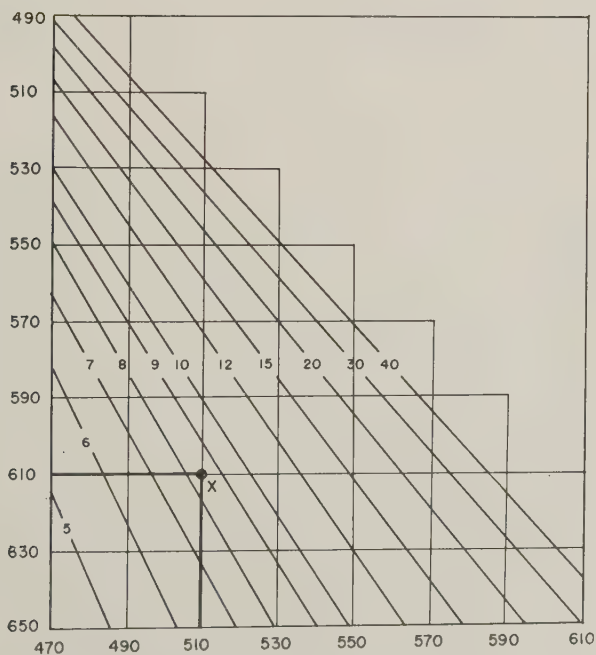


FIG. 4. Dispersion correction chart. Corrections $\times 10^3$. The point X refers to example.

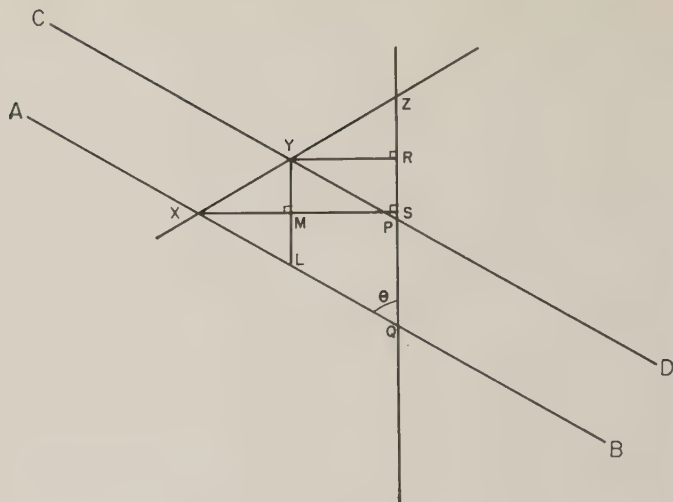


FIG. 5. Geometric proof for Fig. 2. Given: $AB \parallel CD$; PQ equals some constant. Then: $ZQ = ZP + PQ$. (2) $PQ = YL$. (3) $YL/MX = ZQ/XS$. (4) $ZQ = (YL)(XS)/MX = (PQ)(XS)/(XS - YR)$. Thus, for fixed values of XS , YR and PQ , ZQ is independent of the slope angle $90^\circ - \theta$.

CONSTRUCTION AND PROOFS

N_D Correction Chart

The geometric solution is given by the equation:

$$ZQ = (XS)(PQ)/XS - YR$$

Fig. 5

The similarity between the geometric construction and Fig. 1 is readily evident. PQ equals $N_D(2) - N_D(1)$ (0.005 in the following calculations). XS and YR are equivalent to the linear distances on the Hartmann Net between λ_1^* and λ_2^* respectively. ZQ is the correction factor $N_D(M) - N_D(L)$. The relationship between XS , YR and λ_1^* , λ_2^* is given by Cauchy's equation. The numerical values shown in Table 1 will con-

TABLE 1. CONVERSION TABLE FOR DETERMINING WAVE LENGTHS EQUIVALENT TO LINEAR DISTANCES IN GEOMETRIC FORMULAE, MEASURED FROM N_D (589 $m\mu$.)

$m\mu$	656	650	640	630	620	610	600	590	589
cm	-5.00	-4.66	-4.04	-3.36	-2.60	-1.79	-0.95	-0.08	0.00
$m\mu$	580	570	560	550	540	530	520	510	
cm	0.86	1.83	2.87	3.95	5.11	6.32	7.60	8.96	
$m\mu$	500	490	486	480	470				
cm	10.41	11.95	12.93	13.59	15.32				

struct a chart approximately 20 cm.×20 cm., and may be scaled to construct a chart of any size. Solving the equation for various values of ZQ (1, 2, 3, etc.) and plotting gives Fig. 2.

Dispersion Correction Chart

The equation for the solution of the dispersion correction is:

$$AB = FD - (CE)(DC)/(YP) \quad \text{Fig. 6}$$

In Table 1 DC equals 18 cm., and values for YP may be obtained by subtraction. For example, YP for 0.510 $m\mu$ and 0.610 $m\mu$ is $8.96 - (-1.79) = 10.75$ cm. Geometric construction is the same as that of the first chart except that the wave lengths need not be scaled to the Hartmann Net, and may be plotted linearly. Actual plotting will give curved lines for correction factor, but these may be drawn as straight lines (Fig. 4) with a small loss in accuracy.

ERRORS

Although the error involved in one determination is a complex function, experience has shown that the accuracy of the chart is well within the experimental error of the method when using white light and the colored Becke line. It is generally less than 0.0004 for Fig. 2 and less than 0.001 for Fig. 4.

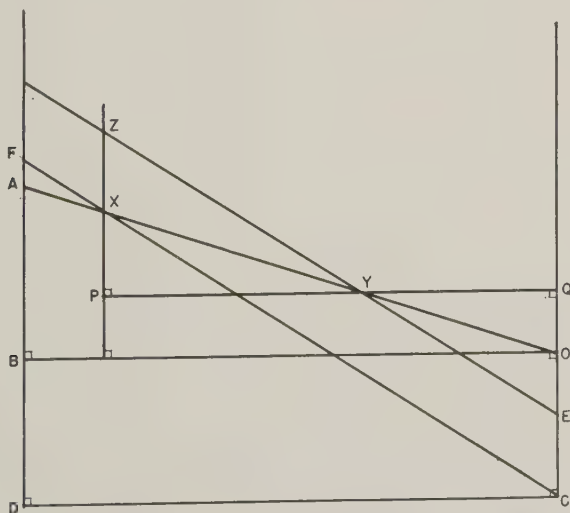


FIG. 6. Geometric proof for Fig. 4. Given: $ZE \parallel FC$ (Note FD is equivalent to $(N_F - N_C)$ of liquids; AB is equivalent to $(N_F - N_C)$ of mineral) Points X and Y represent wave lengths of agreement. (1) $ZP = XP + EC$. (2) $ZP/PY = FD/DC$. (3) $ZP = (PY)(FD)/DC$. (4) $XP = (PY)(FD)/(DC) - (EC)$. (5). $XP/YP = AB/DC$. (6) $AB = (XP)(DC)/YP$. (7) $AB = FD - (CE)(DC)/YP$ by substitution for (XP) .

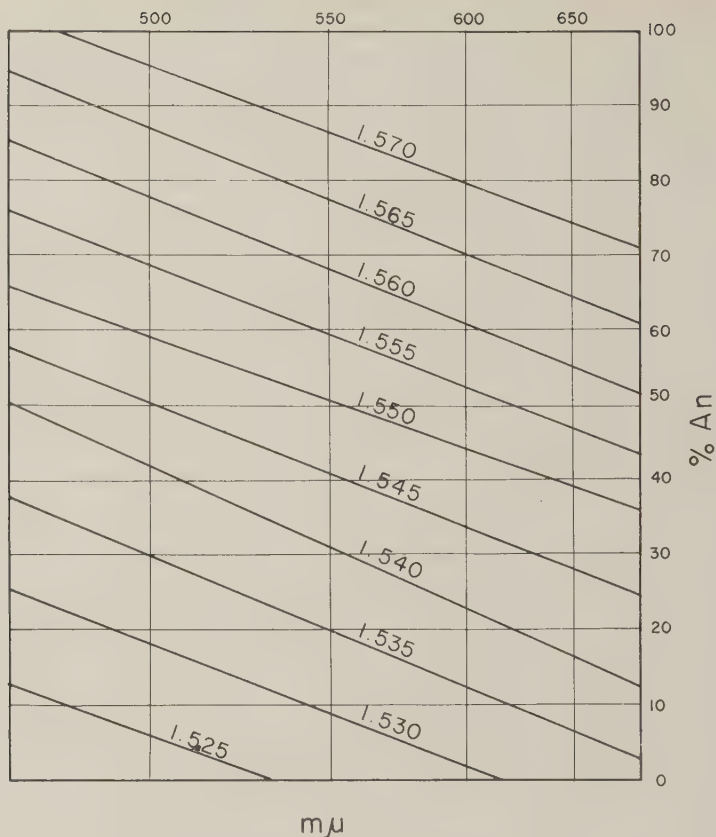


FIG. 7. Plagioclase composition diagrams. The slanting lines represent various possible liquids.

PLAGIOCLASE DIAGRAM

Tsuboi's classic study of the relation of index to composition in feldspars is well known (Tsuboi, 1923, 1934). However, Tsuboi's original diagram was a rather complicated affair involving several steps. A simplification adapted to high dispersion liquids is shown in Fig. 7. After determining the wave length of agreement for a particular liquid, one reads up to the appropriate liquid line and across to the composition. Figure 7 is for N_x' only, but others (N_x , N_z') are just as simply constructed.

CONSTRUCTION

Using Tsuboi's original chart, determine the composition for as many wave lengths per liquid as deemed necessary. Using composition and

wave length of agreement as abscissa and ordinate as desired, plot the points for each liquid on a new graph and connect them with smooth curved lines (the charts shown were plotted using only two points per liquid and have a probable calculated error in composition of less than 2%). It will be noted that the wave lengths are scaled as in a Hartmann Net. This is necessary only if one wishes to have the liquid lines plot as nearly straight lines. Figure 7 was based on University of Wisconsin liquids.

ACKNOWLEDGMENTS

The author wishes to acknowledge his deep gratitude to Professor R. C. Emmons of the University of Wisconsin, without whose encouragement and aid this paper would not have been written. Mr. Paul Ribbe of the University of Wisconsin contributed valuable suggestions concerning the construction of the graphs.

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MINERALOGICAL APPLICATIONS OF ELECTRON DIFFRACTION

II. STUDIES OF SOME VANADIUM MINERALS OF THE COLORADO PLATEAU*

MALCOLM ROSS, *U. S. Geological Survey, Washington 25, D. C.*

ABSTRACT

A method of obtaining precise unit-cell data from oriented electron diffraction powder patterns is described. Partial unit-cell data obtained from such patterns and a description of the crystal habit are given for the following minerals: hewettite ($\text{CaV}_6\text{O}_{16} \cdot 9\text{H}_2\text{O}$), barne-
site ($\text{Na}_2\text{V}_6\text{O}_{16} \cdot 3\text{H}_2\text{O}$), a new hydrated calcium vanadyl vanadate, corvusite ($\text{V}_2\text{O}_4 \cdot 6\text{V}_2\text{O}_5 \cdot n\text{H}_2\text{O}$), a "corvusite-like" mineral ($\text{V}_2\text{O}_4 \cdot 5\text{V}_2\text{O}_5 \cdot x\text{H}_2\text{O}$), fernandinite ($\text{CaO} \cdot \text{V}_2\text{O}_4 \cdot 5\text{V}_2\text{O}_5 \cdot 14\text{H}_2\text{O}$), navajoite ($\text{V}_2\text{O}_5 \cdot 3\text{H}_2\text{O}$), a "corvusite-like" mineral associated with navajoite, steigerite ($\text{Al}_2\text{O}_3 \cdot \text{V}_2\text{O}_5 \cdot 6\frac{1}{2}\text{H}_2\text{O}$), fervanite ($\text{Fe}_4\text{V}_4\text{O}_{16} \cdot 5\text{H}_2\text{O}$), and simplotite ($\text{CaV}_4\text{O}_9 \cdot 5\text{H}_2\text{O}$). For many of these minerals the unit-cell data are new. Structural relations between several of these minerals are indicated.

INTRODUCTION

A large number of fine-grained minerals for which there are no unit-cell data have been described in the literature. Such minerals, particularly those which have layer structures, are not well suited for single crystal x -ray diffraction analysis. On the other hand, the electron diffraction method is better suited for analysis of many fine-grained minerals, particularly those which possess perfect cleavage.

In this paper (Part II) the techniques of interpreting single crystal electron diffraction patterns described elsewhere (Ross and Christ, 1958, hereafter referred to as Part I) have been used in the determination of partial unit-cell data for fine-grained vanadium minerals from the Colorado Plateau. A description of the crystal habit as revealed by electron micrographs is included.

I am especially indebted to Alice D. Weeks for providing a large number of the mineral specimens studied and for her valuable advice concerning their mineralogy. I should also like to thank C. L. Christ and E. J. Dwornik for their advice on various problems, M. E. Thompson for permission to publish data on a new calcium vanadate, and Richard L. Hunt for assistance with some of the experimental work. This work is part of a program being conducted by the U. S. Geological Survey on behalf of the Division of Research of the U. S. Atomic Energy Commission.

THE INTERPRETATION OF ELECTRON DIFFRACTION PATTERNS OF ORIENTED POLYCRYSTALLINE MATERIALS

Crystals which possess perfect cleavage tend to orient on the specimen mount so that the cleavage face is parallel to the substrate and with all

* Publication authorized by the Director, U. S. Geological Survey.

possible orientations of the crystals around the normal to the cleavage face. The type of crystallographic data that can be obtained from the electron diffraction powder patterns of thin crystals oriented in this way will depend upon the plane of orientation and the crystal system of the material in question (Part I, Table 4). For example, powder patterns of thin colemanite crystals oriented on (010) will give directly d_{100} and d_{001} whereas powder patterns of thin KClO_3 crystals oriented on (001) will give directly a and b . The theory appropriate to these conclusions is given in Part I.

As the crystals contributing to an oriented powder pattern are very small, the intensity of the diffracted beam will decrease appreciably with slight deviations of the crystal face from the horizontal. As a result the rings are usually sharp and ideal for precise measurement. If a means can be found to index the powder patterns of oriented materials, measurement of the patterns will yield much more accurate unit-cell data than those obtained from the measurement of spot patterns.

Indexing the oriented powder pattern is fairly simple if a photograph can be obtained showing a single crystal pattern superimposed on rings. This type of pattern is readily obtained when a large crystal and a large number of small crystals, oriented randomly in the plane of the substrate, lie in the path of the electron beam. An example of this type of diffraction pattern may be seen in Fig. 2. An oriented powder pattern will, in general, show rings (usually weak) that cannot be indexed on the basis of the superimposed spot pattern. Such "extra" rings usually appear because additional reciprocal lattice rods are brought into the sphere of reflection by the incompletely oriented crystals.

PREPARATION AND EXAMINATION OF SAMPLES

The vanadium minerals examined in this study have the following characteristics: fine-grained nature; platy, fibrous, or lathlike habit; perfect pinacoidal cleavage and probable layer structure. All the minerals studied tend to orient with a cleavage face parallel to the specimen mount, the "thin direction" of the crystal being parallel to the electron beam. The samples were prepared for examination by grinding lightly in distilled water, placing a droplet of the dispersed material on a collodion mount, and then drying in air. Electron micrographs and SAD (Selected Area Diffraction) spot patterns were usually obtained prior to examination with the electron diffraction unit. External aluminum or β -tin standards were used for most EDU (Electron Diffraction Unit) patterns. A description of the crystal habit as revealed by electron micrographs is given for each mineral. The crystallographic orientation of the minerals is obtained by comparing the relative orientations of the electron micrograph and SAD spot patterns of a single crystal.

The high vacuum (about 10^{-5} cm. Hg) in the electron microscope and diffraction unit offers a severe environment for the analysis of many minerals. The heating effect of the electron beam at its highest intensity (at cross-over) can be tremendous, fusing such stable materials as tourmaline (Barnes, 1950, p. 410) and TiO_2 (Watson, 1948, p. 718). This heating effect is dependent upon the thickness and density of the sample. Dense, thick particles of a given material are much more readily fused than thin flakes or plates of the same material. Few materials are stable enough to resist alteration or destruction when the beam is at the highest intensity. On the other hand, very unstable materials such as gypsum have been successfully examined at the lower beam intensities with the electron microscope or diffraction unit. The assumption that these instruments are "furnaces" is therefore misleading, for a skilled operator can prepare and examine specimens of many unstable materials in essentially their original state.

Minerals that contain interstitial water between the structural sheets are dehydrated to some extent by the vacuum of the electron microscope and electron diffraction unit. This water is very loosely held by such minerals. Although dehydration does not alter the basic layer structure, it does, commonly, change the dimensions of the crystal in a direction normal to the sheets. Minerals that readily lost interstitial water include the montmorillonite clay minerals, autunite $[\text{Ca}(\text{UO}_2)_2(\text{PO}_4)_2 \cdot 10\text{--}12\text{H}_2\text{O}]$, tyuyamunite $[\text{Ca}(\text{UO}_2)_2\text{V}_2\text{O}_8 \cdot 10\text{H}_2\text{O}]$, carnotite $[\text{K}_2(\text{UO}_2)_2\text{V}_2\text{O}_8 \cdot 3\text{H}_2\text{O}]$, and hewettite $(\text{CaV}_6\text{O}_{16} \cdot 9\text{H}_2\text{O})$. Most of the minerals examined in this study contain interstitial water.

Minerals that contain water as an integral part of the structure will, upon dehydration, show a complete change in structure. Such transformations have often been observed in the electron microscope and diffraction unit. For example, gypsum has been observed inverting to anhydrate as the intensity of the electron beam is increased.

EVALUATION OF UNIT-CELL CONSTANTS FROM EDU POWDER PATTERNS

EDU spot patterns superimposed on rings were obtained from all minerals examined in this study. The rings were indexed on the basis of these composite patterns, the sharpest higher order rings being measured to give the unit-cell data. The relation

$$d_x \simeq D_s d_s / D_x$$

was used to calculate the approximate direct lattice spacings (within 0.2 per cent). If the precision of measurement is less than 0.2 per cent the approximate values of d_x are corrected by multiplying by the proper

value of the factor $(\cos 2\theta_s)/(\cos 2\theta_x \cos \theta_s)$. The value of this factor for each direct lattice spacing is found on the graph shown in Part I, Fig. 9.

A measuring microscope was used to measure the powder patterns. When more than one powder pattern was used to determine the unit-cell data, the precision of the determination is given. This value includes the maximum and minimum values of the lattice constant as obtained from several measurements.

As the spot and powder patterns give only two dimensions of the reciprocal lattice, no assignment to a crystal system is made. The following conventions are used in designating the unit-cell constants unless some other crystallographic orientation has been assigned from single crystal *x*-ray study:

- 1) If the spot pattern shows a plane of symmetry, the pattern is assumed to represent a projection of the $hk0$ reciprocal net. The largest cell constant is taken as a . By symmetry $\gamma=90^\circ$.
- 2) If the spot pattern shows only a center of symmetry, the pattern is assumed to represent a projection of the $h0l$ reciprocal net. A primitive cell with the smallest β angle is chosen (unless the presence of "forbidden reflections" indicate otherwise) with a $a > c$.

An attempt is made to predict the systematic absences of reflections by examination of the spot patterns of crystals of varying thicknesses.

MINERALOGIC DATA

Hewettite and barnesite

Hewettite and metahewettite, of the same composition except for water content ($\text{CaV}_6\text{O}_{16} \cdot n\text{H}_2\text{O}$), were first described by Hillebrand, Merwin, and Wright (1914). These minerals are found as nodular aggregates and coatings of fibers, elongate [010], and are deep red in color. Hewettite readily loses water and changes from the $9\text{H}_2\text{O}$ hydrate to a lower hydrate with $3\text{H}_2\text{O}$. Barnes (1955) made dehydration studies of the hewettite minerals and gave evidence that hewettite and metahewettite are structurally identical. He finds that hewettite exists in at least three hydrated forms with $3\text{H}_2\text{O}$, $6\text{H}_2\text{O}$ (probably), and $9\text{H}_2\text{O}$. Barnes (1955, p. 690) suggests that the name metahewettite is unnecessary unless it be retained to designate one of the hydrate phases. Alice D. Weeks suggests (personal communication) that the name metahewettite be used to designate the common $3\text{H}_2\text{O}$ hydrate phase.

Electron micrographs and SAD patterns were made from an analyzed hewettite collected by Nancy G. Ryan from the Hummer mine, Jo Dandy Group, Montrose County, Colorado. The electron micrographs

show that the crystals are well developed laths elongated [010], and flattened {001}. An x-ray powder pattern of the material showed that only hewettite was present. Measurements of six EDU powder patterns of the hewettite from the Hummer mine give the following results: $a = 12.23 \pm 0.03 \text{ \AA}$ and $b = 3.605 \pm 0.005 \text{ \AA}$ ($\gamma = 90^\circ$).

A second sample of hewettite collected by B. N. Webber and L. B. Riley from the Cactus Rat mine, Thompson district, Grand County, Utah, gives electron micrographs and electron diffraction patterns identical to those of the Hummer mine material. The spectrographic analysis of a 10-mg. split of the sample used for this study shows: $>10\%$ V; $1\text{--}10\%$ Ca, Si; $0.1\text{--}1\%$ Na, Al, Fe (Katherine Valentine, analyst). An x-ray powder pattern, however, indicates that a small amount of barnesite is in the sample. Measurement of six EDU powder patterns of this material gives the following results: $a = 12.22 \pm 0.02 \text{ \AA}$ and $b = 3.605 \pm 0.003 \text{ \AA}$ ($\gamma = 90^\circ$).

Figure 1 shows a typical hewettite SAD spot pattern; Fig. 2 shows an EDU pattern of hewettite with spots superimposed on rings.

Barnesite, $\text{Na}_2\text{V}_6\text{O}_{16} \cdot 3\text{H}_2\text{O}$, a sodium analog of hewettite described by Alice D. Weeks, M. E. Thompson, and A. M. Sherwood (written communication) has the same habit as hewettite and is deep red changing

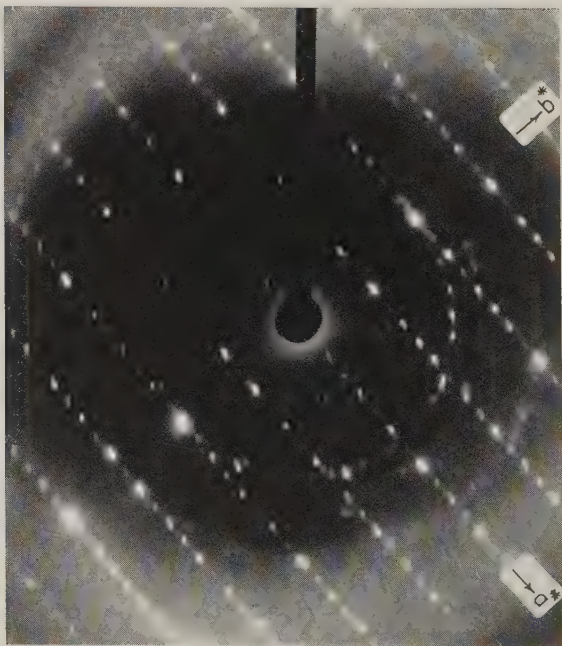


FIG. 1. SAD spot pattern of hewettite.

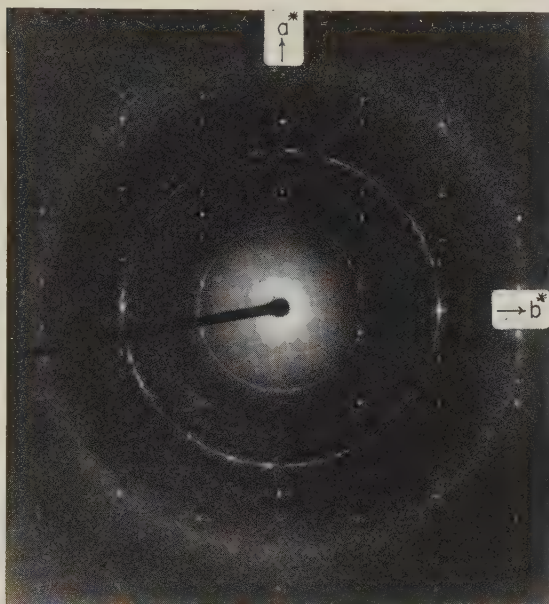


FIG. 2. EDU pattern of hewettite showing spots superimposed on rings.

to brownish-red on exposure to air. Barnes and Qurashi (1952), and Barnes (1955) by *x*-ray analysis found barnesite from Cactus Rat mine, Grand County, Utah (Harvard no. 98019), to be monoclinic with $a = 12.18 \pm 0.02 \text{ \AA}$, $b = 3.614 \pm 0.005 \text{ \AA}$, $c = 7.80 \pm 0.03 \text{ \AA}$ and $\beta = 95^\circ 0' \pm 20'$.

Electron micrographs and electron diffraction patterns of barnesite collected by Alice D. Weeks from the Cactus Rat mine, Thompson district, Grand County, Utah, are nearly identical to those of hewettite. Measurement of six EDU powder patterns gives a cell with $a = 12.17 \pm 0.02 \text{ \AA}$ and $b = 3.604 \pm 0.005 \text{ \AA}$ ($\gamma = 90^\circ$).

The similarity between the spacings and intensities of the spot patterns of hewettite and barnesite indicates that the two minerals have similar structures as projected on the *a*-*b* plane. The results of the hewettite and barnesite study is summarized in Table 1.

A new hydrated calcium vanadyl vanadate

A mineral collected by D. P. Elston from the J. J. mine, Bull Canyon mining district, Montrose County, Colorado, and to be described by M. E. Thompson appears to be a new calcium vanadyl vanadate. Independently, Alice D. Weeks (personal communication, 1956) found a mineral from the East Carrizo mines, San Juan County, New Mexico, which gives the same characteristic *x*-ray powder pattern as Miss Thompson's

TABLE 1. ELECTRON DIFFRACTION AND X-RAY UNIT-CELL DATA FOR HEWETTITE AND BARNESITE

Hewettite	Hewettite ¹	Barnesite
Hummer mine	Cactus Rat mine	Cactus Rat mine
$\text{CaV}_6\text{O}_{16} \cdot n\text{H}_2\text{O}$	$\text{CaV}_6\text{O}_{16} \cdot n\text{H}_2\text{O}$	$\text{Na}_2\text{V}_6\text{O}_{16} \cdot 3\text{H}_2\text{O}$
Electron diffraction data	Electron diffraction data	Electron diffraction data
$a = 12.23 \pm 0.03 \text{ \AA}$	$a = 12.22 \pm 0.02 \text{ \AA}$	$a = 12.17 \pm 0.02 \text{ \AA}$
$b = 3.605 \pm 0.005 \text{ \AA}$	$b = 3.605 \pm 0.003 \text{ \AA}$	$b = 3.604 \pm 0.005 \text{ \AA}$
$\gamma = 90^\circ$	$\gamma = 90^\circ$	$\gamma = 90^\circ$
		x-ray data ²
		monoclinic
		$a = 12.18 \pm 0.02 \text{ \AA}$
		$b = 3.614 \pm 0.005 \text{ \AA}$
		$c = 7.80 \pm 0.03 \text{ \AA}$
		$\beta = 95^\circ 0' \pm 20'$
		Space group: $P2-(C_2')$ (probable)

¹ A small amount of barnesite is present in the sample.

² Barnes and Qurashi (1952) and Barnes (1955).

material. The chemical analyses of the vanadates from the East Carrizo mines and the J. J. mine are essentially identical.

Electron micrographs show the sample from the J. J. mine is composed of thin angular flakes and well-developed, sometimes terminated, laths elongated [010] and flattened {001}. Figure 3 shows an electron micrograph of one of the terminated crystals. Measurement of three EDU powder patterns gives the following results: $a = 12.18 \pm 0.01 \text{ \AA}$ and $b = 3.627 \pm 0.003 \text{ \AA}$ ($\gamma = 90^\circ$). The intensities of the spots are very close to those of the hewettite patterns. One SAD spot pattern (Fig. 4) shows very weak spots which appear to triple the cell in the b direction. As this was the only pattern of many examined that shows such spots it is difficult to decide on their significance.

Electron micrographs (Fig. 5) show that the sample from the East Carrizo mines is composed of irregularly shaped flakes and long, narrow laths or fibers elongated [010] and flattened {001}. Spot patterns of this sample are essentially identical to the typical spot patterns of the sample from the J. J. mine. Measurement of five EDU powder patterns gives the following unit-cell data: $a = 12.15 \pm 0.01 \text{ \AA}$ and $b = 3.612 \pm 0.004 \text{ \AA}$ ($\gamma = 90^\circ$). The results of the study on this new calcium vanadyl vanadate are given in Table 2.

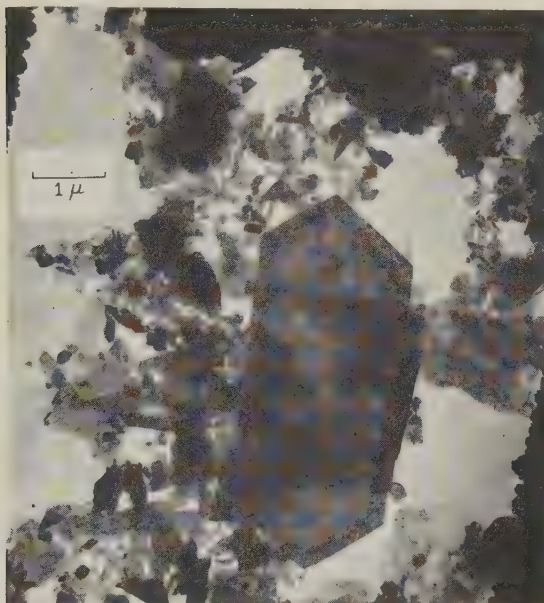


FIG. 3. Electron micrograph of a new hydrated calcium vanadyl vanadate from the J. J. mine.

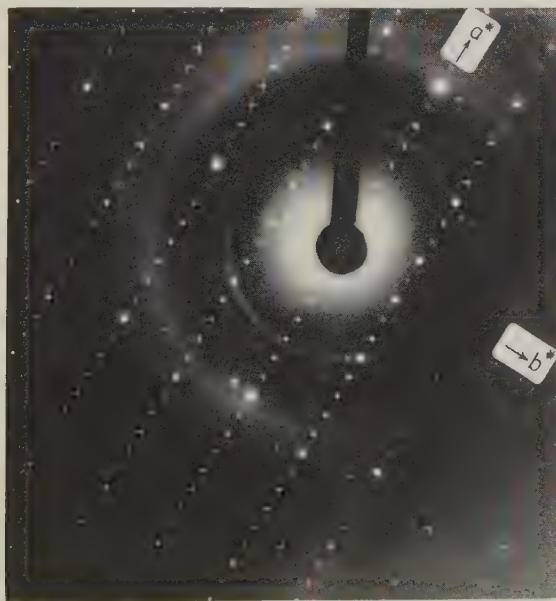


FIG. 4. SAD spot pattern of a new hydrated calcium vanadyl vanadate from the J. J. mine superimposed on the powder pattern of the internal aluminum standard.

TABLE 2. ELECTRON DIFFRACTION UNIT-CELL DATA FOR TWO SAMPLES OF A NEW CALCIUM VANADYL VANADATE

J. J. mine ¹	East Carrizo mines
$a = 12.18 \pm 0.01 \text{ \AA}$	$a = 12.15 \pm 0.01 \text{ \AA}$
$b = 3.627 \pm 0.003 \text{ \AA}$	$b = 3.612 \pm 0.004 \text{ \AA}$
$\gamma = 90^\circ$	$\gamma = 90^\circ$

¹ One SAD spot pattern showed weak spots which appear to make $b = 3 \times 3.627 \text{ \AA}$.

The similarity between the spacings and intensities of the spot patterns of both samples of this new vanadate to those of hewettite indicates that the two minerals have similar structures as projected on the a - b plane.

Fernandinite, corvusite and the "corvusite-like" minerals

A number of vanadium (IV, V) minerals have been loosely described as corvusite or "corvusite-like" and are characterized by their massive or sometimes fibrous appearance and dark-brown, blue-black or dark-green color. Fernandinite is probably closely related to this group of minerals.



Fig. 5. Electron micrograph of a new hydrated calcium vanadyl vanadate from the East Congo mines.

Schaller (1915) described fernandinite from Minasragra, Peru, as a massive dull green mineral with a probable formula of $\text{CaO} \cdot \text{V}_2\text{O}_4 \cdot 5\text{V}_2\text{O}_5 \cdot 14\text{H}_2\text{O}$.

Electron micrographs of the type fernandinite from Minasragra, Peru, (USNM R-5706) disclose angular plates or flakes flattened $\{001\}$. Measurement of four EDU powder patterns gives a cell with $a = 11.69 \pm 0.04$ Å and $b = 3.674 \pm 0.004$ Å ($\gamma = 90^\circ$). The spot patterns of fernandinite show only those reflections for which $h+k=2n$. Due to the thinness of the fernandinite crystals "forbidden reflections" (where $h+k=2n+1$) should appear because of projection of reciprocal lattice rods of the upper levels ($l=1, 2, 3$, etc.) into the sphere of reflection. The fact that "forbidden reflections" do not appear probably indicates that reciprocal lattice points for which $h+k=2n+1$ are missing in the upper l levels. If such reciprocal lattice points are missing, fernandinite has a C -centered cell.

Henderson and Hess (1933) proposed the formula $\text{V}_2\text{O}_4 \cdot 6\text{V}_2\text{O}_5 \cdot x\text{H}_2\text{O}$ for corvusite. Their original material was a brown massive sample (USNM 96807) and a bluish-black massive sample (USNM 96806).

Electron micrographs of the brown corvusite (USNM 96807) show that it is a poorly crystallized, very fine-grained material with an average particle size of less than 0.2 microns. The individual particles are irregularly shaped flakes. Spot patterns show only those reflections for which $h+k=2n$ indicating a probable C -centered cell. Measurement of six EDU powder patterns gives the following unit cell data: $a = 11.6 \pm 0.1$ Å and $b = 3.65 \pm 0.01$ Å ($\gamma = 90^\circ$).

Weeks and Thompson (1954) studied "corvusite-like" minerals from several localities and found them commonly massive and blue-black to greenish black in color. An excellent sample of a "corvusite-like" mineral studied by Weeks and Thompson (personal communication) was collected by Daniel Milton from Incline No. 2, Monument No. 2 mine, Apache County, Arizona. It is a fibrous blue-black material with a probable formula of $\text{V}_2\text{O}_4 \cdot 5\text{V}_2\text{O}_5 \cdot x\text{H}_2\text{O}$. Electron micrographs show that the crystals of this mineral are long, well-developed laths, elongated $[010]$ and flattened $\{001\}$, similar to well-crystallized hewettite. Spot patterns show only those reflections for which $h+k=2n$ indicating a probable C -centered cell. Measurement of six EDU powder patterns gives the following results: $a = 11.69 \pm 0.03$ Å and $b = 3.662 \pm 0.008$ Å ($\gamma = 90^\circ$). Figure 6 shows an EDU pattern of the Monument No. 2 material having spots superimposed on rings.

The similarity between the spacings and intensities of the spot patterns of fernandinite, corvusite and the "corvusite-like" mineral indicates that they have similar structures as projected on the a - b plane.

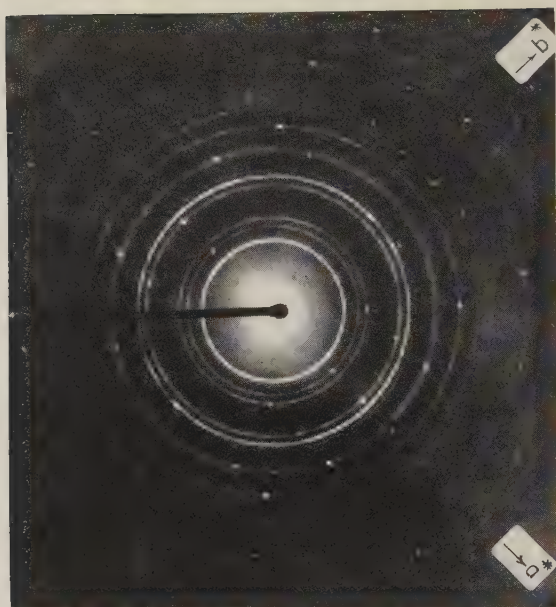


FIG. 6. EDU pattern of a "corvusite-like" mineral from the Monument No. 2 mine, showing spots superimposed on rings.

Weeks and Thompson (1954, p. 47) have noted that the x -ray powder patterns of fernandinite and corvusite are similar. The chemical analyses and diffraction data for these minerals are given in Table 3.

Navajoite

Navajoite, described by Weeks, Thompson and Sherwood (1955) as a hydrated vanadium pentoxide, is a dark brown, soft, fibrous mineral similar to hewettite in habit.

Chemical analysis of navajoite by A. M. Sherwood shows:

V_2O_5	71.68%
V_2O_4	3.08
Fe_2O_3	3.58
SiO_2	1.20
CaO	0.22
H_2O	20.30
<hr/>	
Total	100.06

They give the formula $V_2O_5 \cdot 3H_2O$ for navajoite and suggest that the CaO can be accounted for by 3 per cent hewettite, SiO_2 by quartz, and Fe_2O_3 by substitution of iron for vanadium in the navajoite structure. Weeks and others indexed an x -ray photograph of the zero layer ($k=0$)

TABLE 3. CHEMICAL ANALYSES AND ELECTRON DIFFRACTION UNIT-CELL DATA FOR FERNANDINITE, CORVUSITE AND A "CORVUSITE-LIKE" MINERAL

Brown corvusite ¹ (USNM 96807)		Blue-black "corvusite-like" mineral ² Monument No. 2 mine, Arizona		Fernandinite ³ Minasragra, Peru	
$V_2O_4 \cdot 6V_2O_5 \cdot xH_2O$		$V_2O_4 \cdot 5V_2O_5 \cdot xH_2O$		$CaO \cdot V_2O_4 \cdot 5V_2O_5 \cdot 14H_2O$	
CaO	0.40	CaO	2.5	CaO	3.83
Fe ₂ O ₃	12.20	Fe ₂ O ₃	2.3	Fe ₂ O ₃	0.90
V ₂ O ₄	7.62	V ₂ O ₄	13.3	V ₂ O ₄	11.63
V ₂ O ₅	50.68	V ₂ O ₅	68.1	V ₂ O ₅	63.33
H ₂ O	15.83	H ₂ O	11.4	H ₂ O	18.07
K ₂ O	2.15	K ₂ O, Na ₂ O, Al ₂ O ₃	1.2	K ₂ O	0.59
Na ₂ O	1.24	SiO ₂ +insol	0.7	MgO	0.07
MgO	2.07		—	MoO ₃	1.58
MoO ₃	0.28	Total	99.5		—
UO ₃	2.94	H ₂ O(-)	6.8	Total	100.00
Sol. SiO ₂	4.21				
Total	99.62				
$a = 11.6 \pm 0.1 \text{ \AA}$		$a = 11.69 \pm 0.03 \text{ \AA}$		$a = 11.69 \pm 0.04 \text{ \AA}$	
$b = 3.65 \pm .01 \text{ \AA}$		$b = 3.662 \pm 0.008 \text{ \AA}$		$b = 3.674 \pm 0.004 \text{ \AA}$	
$\gamma = 90^\circ$		$\gamma = 90^\circ$		$\gamma = 90^\circ$	
$hkl: h+k=2n$		$hkl: h+k=2n$		$hkl: h+k=2n$	

¹ Henderson and Hess (1933, p. 201). E. P. Henderson, analyst, recalculated after deducting insoluble.

² Robert Meyrowitz, U. S. Geological Survey, analyst.

³ Palache, Berman, and Frondel (1951, p. 1062). W. T. Schaller, analyst, recalculated to 100 after deducting insoluble.

obtained by rotating a fiber bundle around [010] in a powder camera and found a cell with $a = 17.43 \text{ \AA}$, $b = 3.65 \text{ \AA}$, $c = 12.25 \text{ \AA}$ and $\beta = 97^\circ$. The b dimension was determined from an x-ray photograph of a small bundle of fibers rotated around [010] in a Weissenberg camera.

Electron micrographs (Fig. 7A) of a portion of "navajoite-like" material (Sample A) collected by Mrs. Weeks from the South Rim workings of the Monument No. 2 mine, Apache County, Arizona (the type locality) show characteristic thin well-developed laths and thicker rods or fibers. The laths and fibers are elongated [010] and flattened {001}. The typical EDU spot pattern obtained from Sample A is shown in Fig. 8. From this pattern the strongest rings could be indexed. Measurement of four powder patterns gives the following unit-cell data: $a = 11.86 \pm 0.02 \text{ \AA}$ and $b = 3.644 \pm 0.002 \text{ \AA}$ ($\gamma = 90^\circ$). Reflections appear only when $h+k$ is even indicating a probable C-centered cell. A few spot patterns were

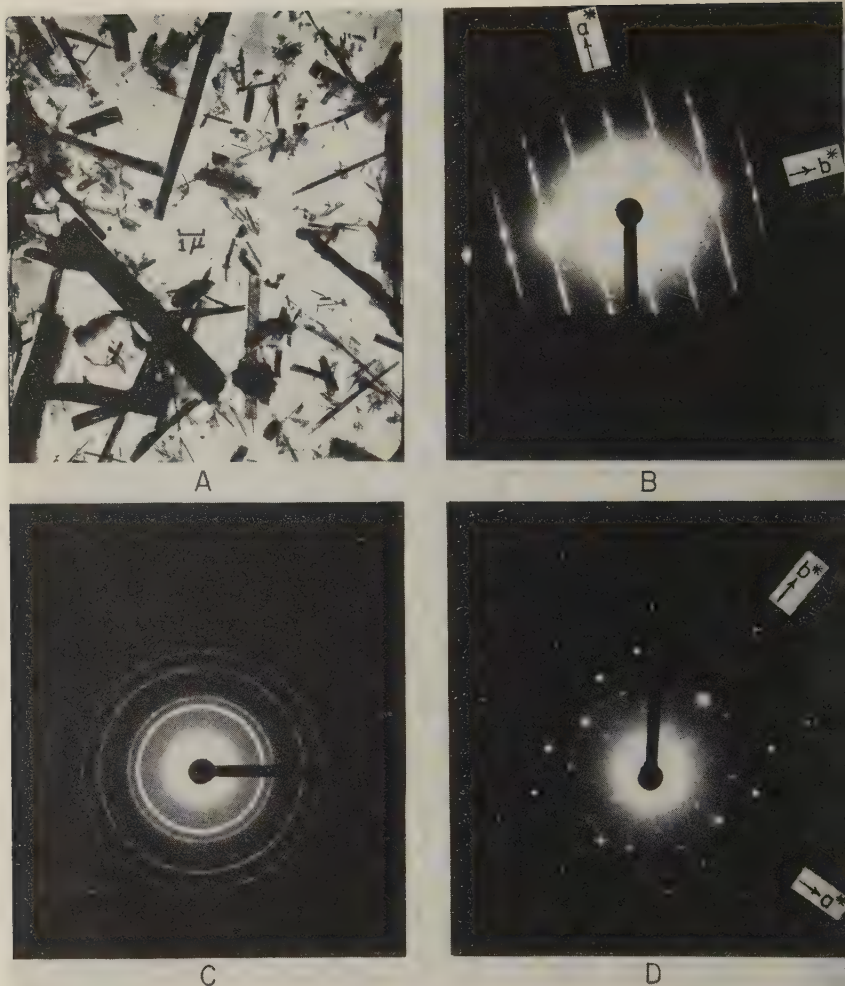


FIG. 7. Electron micrograph and diffraction patterns of navajosite and a "corvusite-like" material associated with navajosite. (A) Electron micrograph of sample A. (B) SAD spot pattern of navajosite. (C) Aluminum standard. (D) SAD spot pattern of a "corvusite-like" mineral associated with navajosite.

obtained from Sample A (Fig. 7B) which are quite different from the "typical" pattern and indicate that a second phase is present in a minor amount. Examination of one of these spot patterns shows a probably C-centered cell ($h+k=2n$) with $a=35 \text{ \AA}$, $b=3.6 \text{ \AA}$ and $\gamma=90^\circ$. The electron diffraction powder patterns of Sample A show several weak reflections which cannot be indexed on the basis of the "typical" spot pattern

but can be on the basis of the "second type" of spot pattern. Also, just outside the strong 020 ring ($d=1.822 \text{ \AA}$) of the major phase is a weak but sharp ring with a spacing of 1.805 \AA which probably represents the 020 reflection of the second phase.

Another sample of "navajoite-like" material (Sample B) was collected by R. G. Coleman from the South Rim workings of the Monument No. 2 mine. Sample B occurred as a seam in sandstone and is dark brown in color and very fibrous. The x -ray powder pattern of Sample B is similar but not identical to the x -ray powder pattern of Sample A.

Electron micrographs of Sample B show that the crystal habit is identical to that of Sample A. Electron diffraction patterns, however, show that the predominant phase in Sample B is identical to the second phase in Sample A. Measurement of seven EDU powder patterns of Sample B, indexed from the pattern shown in Fig. 9, gives the following unit-cell data: $a=34.7 \pm 0.2 \text{ \AA}$ and $b=3.606 \pm 0.004 \text{ \AA}$ ($\gamma=90^\circ$). The hkl reflections appear only when $h+k$ is even. A few spot patterns (Fig. 7D) were obtained from Sample B which were identical to the spot patterns of the major phase in the Sample A. Measurement of one of these spot patterns gives a cell with $a=12 \text{ \AA}$ and $b=3.6 \text{ \AA}$ ($\gamma=90^\circ$). Reflections appear only when $h+k=2n$.

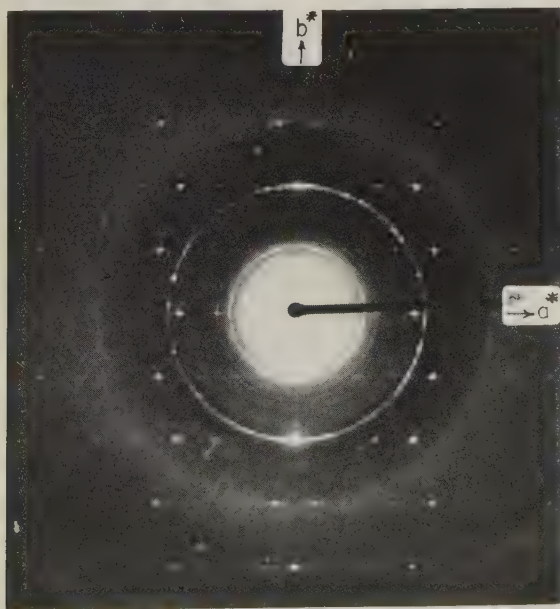


FIG. 8. EDU pattern of a "corvusite-like" mineral associated with navajoite, showing spots superimposed on rings.

The above data indicate that both samples from Monument No. 2 mine contain two phases. The major phase in Sample B is navajoite, whereas the major phase in Sample A appears to be a "corvusite-like" mineral. This apparent "corvusite-like" mineral may represent a stage in the oxidation of corvusite to navajoite, and its presence in navajoite samples may account for the appearance of V^{+4} in the chemical analysis of navajoite. The reason why so little navajoite appeared in Sample A may be due to sampling. Electron diffraction data show that the 17.4 Å reflection is a 200 reflection instead of a 100 reflection as Weeks and others had assumed from x-ray studies. Combination of the electron diffraction data with the x-ray data of Weeks and others would suggest that navajoite is monoclinic, C-centered with the following unit-cell constants: $a=34.8$ Å, $b=3.606$ Å, $c=12.25$ Å, and $\beta=97^\circ$. The results of this study are summarized in Table 4.

Fervanite

Hess and Henderson (1931) describe fervanite from Gypsum Valley, Colorado as a hydrous ferric vanadate with the formula $Fe_4V_4O_{16} \cdot 5H_2O$. The mineral is fibrous, golden brown with a brilliant luster. Extinction is slightly inclined to the length of the fiber with 2V small, suggesting monoclinic symmetry (Hess and Henderson, 1931, p. 274).

Fervanite collected by H. T. Evans from the Hummer mine, Montrose County, Colorado was examined. The material is golden brown and very fibrous, closely resembling hewettite in habit. Electron micrographs (Fig. 10) show that the material is composed of long narrow fibers elongated $\{001\}$ and flattened $\{010\}$. The width of the fibers in the a direction averages about 0.5 microns. Measurement of four EDU powder patterns gives a cell with $a=9.02 \pm 0.01$ Å, $c=6.65 \pm 0.03$ Å and $\beta = 103^\circ 20' \pm 20'$. A primitive cell with the smallest β angle was chosen. Fig. 11 shows an EDU spot pattern of fervanite.

Steigerite

Henderson (1935) describes steigerite as a hydrous aluminum vanadate, $Al_2O_3 \cdot V_2O_5 \cdot 6\frac{1}{2}H_2O$; canary yellow in color with a waxy luster.

Electron micrographs of steigerite from Gypsum Valley, Colorado (USNM E-5108) show that the material is composed of thin, poorly developed laths and angular flakes, many of which are less than one micron in size. The laths and flakes are flattened $\{010\}$. Measurement of seven powder patterns indexed from an EDU spot pattern gives the following unit-cell data: $a=12.92 \pm 0.03$ Å, $c=10.98 \pm 0.04$ Å and $\beta=121^\circ 13' \pm 30'$. A primitive cell with the smallest β angle was chosen.



FIG. 9. EDU pattern of navajoite, showing spots superimposed on rings.



FIG. 10. Electron micrograph of fervanite.

TABLE 4. ELECTRON DIFFRACTION AND X-RAY UNIT-CELL DATA FOR NAVAJOITE AND A "CORVUSITE-LIKE" MINERAL ASSOCIATED WITH NAVAJOITE

Sample A ¹ Electron diffraction data	Sample B ² Electron diffraction data
Major phase ("corvusite-like" mineral) $a = 11.86 \pm 0.02 \text{ \AA}$ $b = 3.644 \pm 0.002 \text{ \AA}$ $\gamma = 90^\circ$ $hkl: h+k=2n$	Minor phase ("corvusite-like" mineral) ⁴ $a = 12 \text{ \AA}$ $b = 3.6 \text{ \AA}$ $\gamma = 90^\circ$ $hkl: h+k=2n$
Minor phase (navajoite) ³ $a = 35 \text{ \AA}$ $b = 3.6 \text{ \AA}$ $\gamma = 90^\circ$ $hkl: h+k=2n$	Major phase (navajoite) $a = 34.7 \pm 0.2 \text{ \AA}$ $b = 3.606 \pm 0.004 \text{ \AA}$ $\gamma = 90^\circ$ $hkl: h+k=2n$
Navajoite (type material) $\text{V}_2\text{O}_5 \cdot 3\text{H}_2\text{O}$ x -ray data ⁵ monoclinic $a = 17.43 \text{ \AA}$ $b = 3.65 \text{ \AA}$ $c = 12.25 \text{ \AA}$ $\beta = 97^\circ$	

¹ "Navajoite-like" material collected by A. D. Weeks from the South Rim workings, Monument No. 2 mine, Apache County, Arizona.

² "Navajoite-like" material collected by R. G. Coleman from the South Rim workings, Monument No. 2 mine, Apache County, Arizona.

³ Cell data obtained from measurement of an EDU spot pattern.

⁴ Cell data obtained from measurement of an SAD spot pattern.

⁵ Weeks, Thompson, and Sherwood (1955, p. 211).

Simplotite

Simplotite, $\text{CaV}_4\text{O}_9 \cdot 5\text{H}_2\text{O}$, described by Thompson, Roach and Meyrowitz (1958) occurs as hemispherical aggregates of dark green, platy crystals. They found by x -ray techniques that the mineral was monoclinic, pseudotetragonal with $a = 8.39 \pm 0.03 \text{ \AA}$, $b = 17.02 \pm 0.05 \text{ \AA}$, $c = 8.37 \pm 0.03 \text{ \AA}$, and $\beta = 90^\circ 25' \pm 5'$.

Electron micrographs of the type material collected by C. H. Roach from the Peanut mine, Montrose County, Colorado, show that it is composed of irregularly shaped plates or flakes varying greatly in size. Measurement of four EDU powder patterns of simplotite gives a cell with $a = c = 8.336 \pm 0.002 \text{ \AA}$ and $\beta = 90^\circ 0' \pm 30'$. An SAD spot pattern of simplotite superimposed on the rings of the aluminum standard is shown in

Thompson, Roach and Meyrowitz (1958, p. 22). The tetragonal symmetry is apparent.

DISCUSSION

The electron diffraction technique can be used to determine unit-cell constants of crystals that may be too fine grained to give useful optical or x-ray data. The importance of the technique for the study of layer

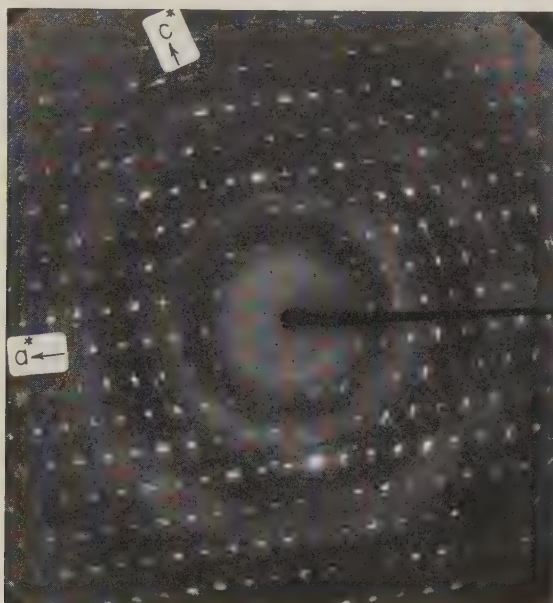


FIG. 11. EDU spot pattern of fervanite.

lattice minerals which are unsuitable for single crystal x-ray diffraction analysis cannot be overstressed, particularly in view of the recent successful applications of electron diffraction methods to the complete solution of crystal structures (Pinsker, 1953; Cowley, 1953*a,b*).

No attempt is made to determine completely the unit-cell of minerals examined in this study either by correlation with x-ray powder patterns or by examination of electron diffraction patterns of "oblique texture."* The use of x-ray powder patterns in connection with the "two-dimensional" electron diffraction data to determine completely the unit-cell of layer lattice minerals presents a number of problems. In general, the

* An "oblique textured" pattern refers to an electron diffraction pattern obtained from a polycrystalline specimen which has preferred orientation, the plane of orientation being inclined from a normal to the electron beam.

x -ray pattern of this type of mineral shows weak and broad or diffuse rings, making measurement difficult. Moreover, indexing of the x -ray powder pattern with electron diffraction until cell data is often ambiguous because a series of closely spaced reflections may appear only as broad bands in the x -ray pattern. Indexing of the x -ray pattern by direct comparison with an index EDU oriented powder pattern is usually impossible because, due to orientation effects, the EDU pattern includes only a restricted number of reflections; whereas the equivalent reflections in the x -ray pattern are usually very weak or missing entirely. The direct lattice spacings given by EDU powder patterns of nonorthogonal crystals may be different from the direct lattice spacings given by the x -ray powder pattern and thus cause further difficulty in indexing. The reasons for the differences in the values of the lattice spacings obtained by the two methods for nonorthogonal crystals are discussed in Part I.

Electron diffraction powder data cannot, in general, be compared to ASTM x -ray data for the identification of fine-grained minerals because of the variations between x -ray and electron diffraction powder patterns mentioned above. In this laboratory a set of standard EDU powder and spot patterns is used to identify unknown materials.

Electron diffraction patterns of "oblique texture" have been used by Pinsker and co-workers to find the unit cell and space group of a number of compounds. Pinsker (1953, p. 93) gives a general account of the interpretation of patterns of "oblique texture." It is beyond the scope of this paper to discuss the use of this type of pattern in evaluation of the unit cell of layer lattice minerals except to point out that in many cases minerals of this type are so thin in one direction that patterns of "oblique texture" give no information about the axial length in the thin direction of the crystal.

Little use was made in this paper of SAD spot or powder patterns for the determination of unit-cell data; however, SAD patterns give the same unit-cell data as do the EDU patterns. Measurements of SAD patterns are not very precise, particularly if the patterns are not internally standardized.

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A RESTUDY OF STEVENSITE AND ALLIED MINERALS*

GEORGE T. FAUST,† JOHN C. HATHAWAY,† AND GEORGES MILLOT‡

ABSTRACT

Stevensite was shown by Faust and Murata to be a montmorillonite group mineral. Brindley subsequently proposed that stevensite consists of an interlayered talc-saponite mineral. A restudy of stevensite shows that it is not an interlayered talc-saponite mineral but rather a type of defect structure. The defect structure arises from a deficiency in the total number of ions in octahedral coordination. This concept of stevensite as a defect structure seems to fit all of the data from x-ray, infrared, differential thermal analysis, and solution studies.

Ghassoulite is shown to contain 0.36% Li_2O and 3.22% F. Experimental studies show it to be like hectorite instead of stevensite.

Synthetic stevensites, the interstratified "saponite-talc" of Alietti, and hanušite are interpreted in terms of the results reported here.

Two new chemical analyses are reported:—Ghassoulite from Djebel Ghassoul, Morocco, and a mixture containing stevensite from Mine Creek, North Carolina.

It is recommended that the names hanušite (=stevensite+pectolite) and ghassoulite (=hectorite) be relegated to the synonymy.

INTRODUCTION

Stevensite was redefined as a member of the montmorillonite group by Faust and Murata (1953) on the basis of differential thermal analysis, staining, x-ray powder diffraction, and chemical studies. Since that time further studies have appeared on stevensite and related minerals and the original literature on hanušite has become available to us. This paper presents the results of further studies and an interpretation of the various data.

During World War II, Kašpar (1942) described the mineral hanušite which he characterized as the magnesium end-member of the supposed pectolite-walkerite series. Fleischer (1943), commenting on the new mineral name, hanušite, suggested the relationship to stevensite. Brindley (1955) made further studies on stevensite and proposed on the basis of his x-ray data that stevensite "may be described as a talc-saponite interlayered mineral."

Millot (1954), unaware of the paper by Faust and Murata, gave the name ghassoulite to a clay from Morocco which is used in laundering. This clay had been described, with an analysis, by Damour (1843). Fleischer (1955) in his discussion of new mineral names, stated that ghassoulite was probably identical with stevensite. Andrea Alietti (1956) described a mineral related to stevensite from Monte Chiaro, Italy, as an interlayered mineral of saponite and talc.

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† U. S. Geological Survey, Washington 25, D. C.

‡ Université de Strasbourg, Strasbourg, France.

DESCRIPTION OF THE SPECIMENS STUDIED

The mineral cabinet of Albert R. Leeds, (the discoverer of stevensite) was bequeathed to the Stevens Institute of Technology at Hoboken, New Jersey, where Leeds had been a professor. In an attempt to locate the type specimen of stevensite, we requested Professor S. C. Williams of that institution to search for it in the Leeds cabinet. Unfortunately, neither the type material nor any specimen of stevensite could be located. Professor Williams did, however, send us a specimen of stevensite which he had obtained for this study from William C. Casperson, at that time curator of minerals in the Paterson Museum, Paterson, New Jersey.

Color references are to Ridgway's standards symbols.

Stevensite No. 1.—Locality: Springfield, New Jersey (U.S.N.M. R4719). Pseudomorphous after pectolite. Pink masses. The powder of the sample used for the chemical analysis and other studies is pale Salmon Color, 9'.OR · O.f. Analysis 1, Table 1.

Geological Occurrence: Late stage hydrothermal replacement of pectolite in cavities in basalt.

Stevensite No. 2.—Locality: Jersey City, New Jersey, (Selfridge No. 51). Deweylite, "Serpentine after pectolite." Color, Wood Brown, 17''' O. Y.

Geological Occurrence: Late stage hydrothermal replacement of pectolite in cavities in basalt. From Professor Paul F. Kerr.

Stevensite No. 3.—Locality: Paterson, New Jersey (F-1104). Pseudomorphous after pectolite. Color of fresh surface is Pale Vinaceous-Fawn 13''' O. Y.-O.f. Color of weathered surface is Vinaceous-Buff, 17''' O. Y.d.

Geological Occurrence: Late stage hydrothermal replacement of pectolite in cavities in basalt. Collected by George T. Faust.

Stevensite No. 4.—Locality: New Street Quarry, Paterson, New Jersey (#685). Pseudomorphous after pectolite. Color of most of the specimen is pale pinkish buff, 17'' O. Y.f. A small portion is safrano pink, 7. R. O. f.

Geological Occurrence: Late stage hydrothermal replacement of pectolite in cavities in basalt. From Professor S. C. Williams.

Stevensite No. 5.—Locality: Mine Creek, Bakersville, Mitchell County, North Carolina (F-48). Insufficient material for naming the color. This sample is impure and contains some serpentine mineral.

Geological Occurrence: Probably a hydrothermal alteration product in dunite. From Dr. C. S. Ross. Analysis 3, Table 1.

Stevensite.—Synthetic. Prepared by N. L. Bowen and O. F. Tuttle (1949). Tentatively identified by them as sepiolite (?). Charge T-2-37. Theoretical Composition of sepiolite-held at 340° C. and at a pressure 30,000 pounds per square inch. Color white.

Ghassoulite.—Locality: Djebel Ghassouel* mine, on the spur of the Moyen Atlas Mountains which dominate the valley of the Moulouya, Ksabi Province, Eastern Morocco. Color—light drab, 17''' O.Y.b. Analysis 4, Table 1.

Geological Occurrence: Tertiary sedimentary argillite. Continental facies. Series include periodically: flint, the "earth ghassoulite," green marl and white limy-marl. Ghassoulite is authigenic in origin in a lake deposit of chemical sediments (Jeannette, 1952; Millot, 1949).

Hectorite.—Locality: About 3 miles south of the village of Hector, San Bernardino County,

* Some transliterations give Rhassoul and Rassoul.

TABLE 1. CHEMICAL ANALYSES OF STEVENSITE, GHASSOULITE, AND HECTORITE

Analysis No.	1	2	3	4	5	6	7	8
Name	Stevensite No. 1	Stevensite No. 2	Stevensite No. 5	Ghassoulite	Ghassoulite	Hectorite	Hanušite	"Talc- saponite"
Locality	Springfield, N. J. U.S.N.M. R4719	Jersey City, N. J. Selfridge No. 51	Mine Creek, Bakersville, N. C. F 48	Djebel Ghas- soul mine Ksabi; Prov- ince, Morocco	Djebel Ghas- soul mine Ksabi; Prov- ince, Morocco	Near Hector, San Bernardino County, Calif.	Liebstadt, Riesengebirge	Monte, Chi- aro, Val di Taro, Italy
SiO ₂	57.30	57.24	47.0	55.02	55.0	55.86	57.37	54.31
Al ₂ O ₃	None	None	—	1.12	1.2	.13	—	1.90
Fe ₂ O ₃	0.32	1.14	—	—	1.4	.03	—	—
FeO	None	0.12	5.5	0.70	—	—	3.57	2.95
MnO	0.21	0.75	—	—	—	—	0.54	n.d.
MgO	27.47	27.89	29.7	24.89	28.0	25.03	18.55	27.44
CaO	0.97	0.38	0.2	0.54	1.01	Tr	4.84	1.01
K ₂ O	0.03	—	—	0.43	0.52	.10	—	0.86
Na ₂ O	0.03	—	—	0.94	—	2.68	—	0.32
Li ₂ O	—	—	—	0.36	—	1.05	—	—
CO ₂	—	—	—	0.30	—	—	—	—
F	None	None	—	3.22	—	5.96	—	—
H ₂ O ⁺	7.17	7.69	17.7	6.42	10.35	2.24	6.49	9.48
H ₂ O ⁻	6.69	4.76	—	7.66	—	9.90	8.68	2.59
Σ	100.19	99.97	100.1	101.68 ¹	97.48 ²	102.98 ³	100.04	100.86
Analyst	K. J. Murata	K. J. Murata	C. Milton	J. J. Fahey	A. A. Damour	R. E. Stevens	J. V. Kašpar	A. Alietti

¹ 101.68-1.36 (0=F)=100.32. Contains 0.08 TiO₂.² Excludes 1.5% sand reported in the original.³ 102.98-2.51 (0=F)=100.47.

California. Color pure white. API clay mineral standard, H-34, used in *x*-ray studies. Analysis and DTA study on other purified samples.

Geological Occurrence: Hydrothermal replacement of tuff enclosed within basaltic lava flows. Analysis 6, Table 1.

Table 1 contains in addition to the analyses cited above the analysis of other minerals pertinent to this discussion. These are analysis 2, stevensite from Jersey City, N. J. (Faust and Murata, 1953); analysis 5, ghassoulite from Morocco (Damour, 1843); analysis 7, hanušite from the Riesengebirge (Kašpar, 1942) and analysis 8, talc-saponite from Italy (Alietti, 1956). Analysis 3, stevensite from North Carolina and analysis 4, ghassoulite from Morocco are new.

DIFFERENTIAL THERMAL ANALYSIS DATA

The differential thermal analyses were made in a modified form of the Alexander, Hendricks, and Nelson (1939) apparatus. The samples were heated in a nickel block at the rate of 12° C. per minute. The records were obtained as photographs. The samples were all air-dried at room temperature and at a relative humidity of about 65%.

DTA patterns of the original sample, stevensite No. 1, studied by Faust and Murata (1953); of stevensite No. 4, from Paterson, New Jersey; of stevensite No. 5, an impure stevensite from Mine Creek, North Carolina; of synthetic stevensite synthesized by Bowen and Tuttle (1949); of ghassoulite from Morocco; and of hectorite from Hector, California are shown in Fig. 1. The general similarity of these patterns is apparent and they are typical of the montmorillonite group of minerals. The sample weights and data obtained from the curves are summarized in Table 2.

The DTA record (C-123) of stevensite No. 1 covers the temperature range of 25° C. (at the beginning of the curve) to 1061° C. (at the end of the curve). At the high temperature end of the curve from about 885° C. to 1061° C. the record is essentially a straight line. It is within this region that talc undergoes its dehydration and its endothermic trough generally has its minimum at about 950° C. to 1000° C. Such a trough does not appear on the curve for stevensite No. 1. The behavior of stevensite No. 4 (C-785), bearing in mind the difference in weight of the samples, is almost the same as that of stevensite No. 1, and the temperatures of the minima of the troughs are very close (see Table 2). The small area under the low-temperature endotherm of the synthetic stevensite (C-519) arises from the drying out of the sample. Drying out has probably been accelerated by the very fine grain size of the synthetic preparation. The *x*-ray powder pattern of this sample shows two weak lines which suggest the presence of a small amount of chrysotile. Stevensite No. 5, (C-546),

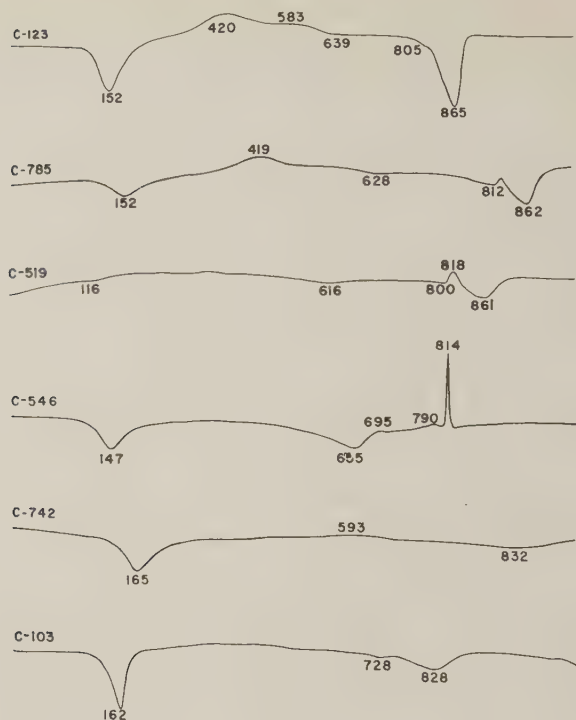


FIG. 1.—Differential thermal analysis curves for Stevensite No. 1, Springfield, New Jersey, C-123; Stevensite No. 4, Paterson, New Jersey, C-785; Synthetic stevensite, Bowen and Tuttle preparation T-2-37, C-519; Stevensite No. 5 with serpentine, Mine Creek, North Carolina, C-546; Ghassoulite, Morocco, C-742; and Hectorite, Hector, California, C-103.

The temperature on the left of the diagram is about 25° C; the temperature at the termination of the curve, on the right, is about 1000° C.

is an impure sample containing a serpentine mineral. The curves of ghassoulite (C-742) and hectorite (C-103) seem most closely alike. It is noteworthy that at the termination of the DTA curve for hectorite a high temperature endothermic trough appears. The onset of this trough is at 1053° C. and as far as this record goes to 1077° C. the minimum was not reached. Such a high temperature endothermic trough in hectorite has been noticed by other investigators, Grim and Rowland (1942) and Schmidt and Heystek (1953).

X-RAY STUDIES

The samples were prepared for x-ray analysis as follows: hectorite was treated with glacial acetic acid diluted with 3 parts of water to remove calcite in the sample, then washed with distilled water. The $<2\mu$ fraction

TABLE 2. OBSERVATIONS ON THE THERMAL ANALYSIS CURVES¹ OF STEVENSITE AND RELATED MINERALS

Mineral	Record no.	Temperature of Troughs and Peaks ° C.						Locality	Weight of sample used (grams)		
		Low temperature		Intermediate temperature		High temperature					
		Endo-thermic	Exo-thermic	Endo-thermic	Exo-thermic	Endo-thermic	Exo-thermic				
Stevensite No. 1	C-123	152	420	639 Inflexion	583	805	865	—	—	Springfield, New Jersey U.S.N.M. R4719	0.4762
Stevensite No. 4	C-785	152	419	628 Inflexion	—	812	862	—	—	Paterson, New Jersey	0.4513
Stevensite Synthetic with a little chrysotile	C-519	116	—	616	—	800	861	—	818	Bowen and Tuttle Prep. No. T-2-37	0.1862
Stevensite No. 5 with ser-pentine	C-546	147	—	655	695	—	—	—	790	814	0.3022
Ghassoulite	C-742	165	—	—	593	—	832	—	—	Morocco	0.3771
Hectorite	C-103	162	—	—	—	728	828	—	—	Hector, California	0.4377

¹ Resistance in the galvanometer circuit for these curves is as follows: 600 ohms for C-123, C-785, C-742, and C-103; 900 ohms for C-519 and 999.9 ohms for C-546.

was obtained by repeated centrifuging and decanting. Stevensite No. 4 was also fractionated by centrifuging but the Stevensite No. 1 and ghas-soulite samples proved too small to handle in this manner. The ghas-soulite probably contains a very small amount of quartz. These (Stevensite No. 1 and ghas-soulite) were dispersed in distilled water in test tubes by a motor driven brush and allowed to stand until most of the coarser particles had settled. Oriented aggregates of the fine fraction of these four minerals were prepared by pipetting some of the clay suspension onto glass slides or by centrifuging the material onto porous plates by the method of Kinter and Diamond (1956). Randomly oriented powders were prepared by drying the clay at room temperature, crushing to a moderately fine powder and sieving the material with a 200 mesh sieve into aluminum mounts with openings 10×41 mm. Care was taken not to pack the powder tightly in order to avoid introducing too much preferred orientation. X-ray patterns were prepared using a Norelco diffractometer. The operating constants are given below:

	<i>For randomly oriented powders</i>	<i>For oriented aggregates</i>
Radiation	Cu, Ni filtered	Cu, Ni filtered
KVP	40	40
MA	20	20
Divergence slit	4°	1°
Receiving slit	.006 inch	.006 inch
Scatter slit	4°	1°
Sample length	41 mm.	44 mm.
Scale factor	8	16-64
Multiplier	1	1
Time constant	4 sec.	4-2 sec.
Scanning speed	2°/min.	2°/min. + 1°/min.
Chart speed	$\frac{1}{2}$ inch/min.	$\frac{1}{2}$ inch/min.

Traces of the patterns obtained from the randomly oriented powders are shown in Fig. 2. There is a close similarity of the patterns, except for the 00 l diffraction maxima and those due to impurities. The characteristically broad asymmetrical hk bands of the montmorillonite group minerals are present in all four samples.

The differences in these minerals are brought out more clearly in the patterns for the untreated oriented aggregates shown in Fig. 3.

The summary of the diffraction data shown by these two groups of patterns is given in Table 3.

The intensities shown in Table 3 are derived from measurements of the relative areas under the diffraction peaks. Relative intensities for 00 l reflections were based on patterns of oriented aggregates whereas the intensities for the hk reflection were based on the patterns of randomly

oriented powders. The strongest reflection in each of these two groups was assigned an intensity of 10. The reasons for assigning these intensities separately are as follows:

(1) The $00l$ reflections may vary considerably in intensity with respect to the hk reflections as a result of preferred orientation.

(2) The wide (4°) slits used with the randomly oriented powder to obtain adequate intensity of the diffracted radiation prevent the recording of diffraction effects below about $5^\circ 2\theta$, and the long spacings of air-dried stevensite occur in this region.

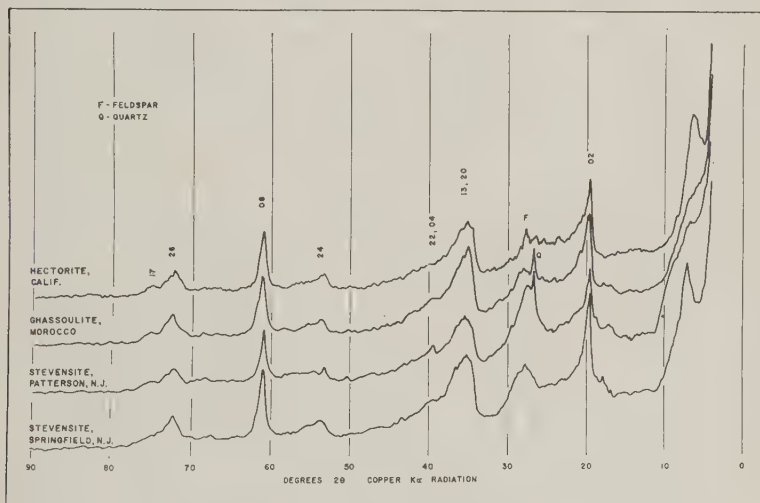


FIG. 2.—X-ray diffractometer patterns of randomly oriented powders of hectorite, ghassoulite, and stevensite.

(3) In spite of the use of a long (41 mm.) sample mount, the entire divergent x-ray beam is not intercepted by the sample below about $20^\circ 2\theta$. Thus lines recorded below $20^\circ 2\theta$ are not strictly comparable with those recorded above this angle. On the other hand, the narrower (1°) slits used with the oriented aggregates and the long sample (44 mm) allow direct comparison of intensities above about $8^\circ 2\theta$. Discrepancies between samples may arise, however, depending on whether Lorentz and polarization factors for random powders, extended faces of single crystals, or some intermediate condition apply to the $00l$ intensities observed in oriented aggregates. The justification of applying strictly any one factor to the correction of such intensities is questionable.

The indices assigned are according to Brindley (1955, p. 241). The long spacing of $\sim 25 \text{ \AA}$ noted by Brindley was found in both Stevensite

TABLE 3. X-RAY POWDER DIFFRACTION DATA FOR HECTORITE, GHASSOULITE, AND STEVENSITE DRIED AT ROOM TEMPERATURE

Mineral	Hectorite			Ghassoulite			Stevensite No. 4			Stevensite No. 1		
Locality	Hector, Calif.			Morocco			Paterson, N. J.			Springfield, N. J.		
Indices	d (Å)	I*	Notes	d (Å)	I*	Notes	d (Å)	I*	Notes	d (Å)	I*	Notes
~001	12.7 - 14.0	10*	variable	12.5 - 14.5	10*	variable	24.5	1*	variable	26.5	2*	variable
~002							12.1 - 13.0	10*	variable	12.5 - 13.0	10*	variable
02, 11		6		4.85 - 5.1	<1	variable	4.78 - 5.2	<1*	variable	4.94 - 5.0	<1*	variable
~003	4.5			4.49	6		4.54	6		4.54	6	
13, 20	3.16 - 3.35	2*	variable	3.18 - 3.24	2*	variable	3.20 - 3.24	3*	variable	3.17 - 3.20	3*	variable
04, 22	2.56	10	very broad	2.56	10	very broad	2.54	10	very broad	2.55	10	very broad
15, 24, 31	2.27	1		2.28	1		2.28	1		2.28	1	
06, 33	1.715	2		1.710	1		1.720	1		1.710	1	
26, 40	1.520	4	sharp	1.516	4	sharp	1.520	4	sharp	1.520	4	sharp
17, 35, 42	1.306	3		1.303	3		1.307	2		1.306	2	
	1.262	1		1.262	1		1.265	1		1.265	<1	

* Intensities of 00l reflections based on oriented aggregates.

Intensities of hk reflections based on randomly oriented powders.

Strongest 00l and hk reflections each assigned an intensity of 10.

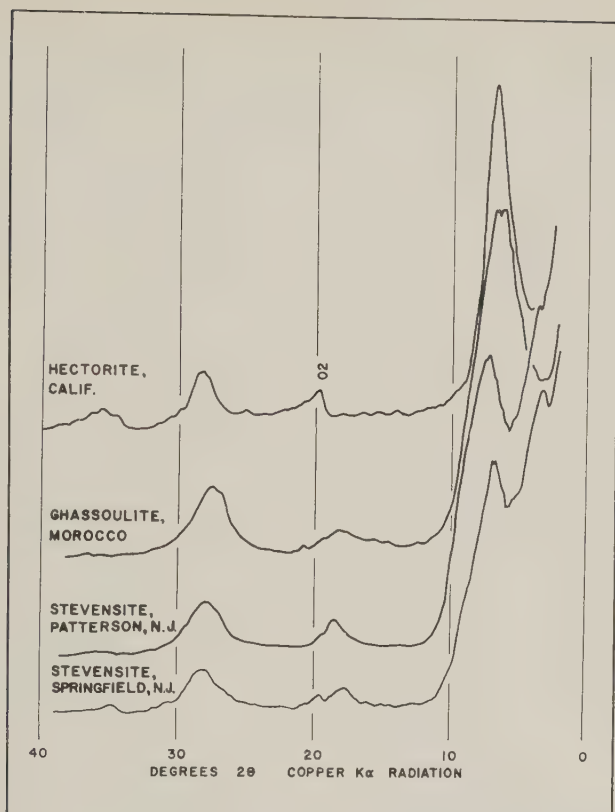


FIG. 3.—X-ray diffractometer patterns of untreated oriented aggregates of hectorite, ghassoulite and stevensite.

No. 4 and Stevensite No. 1. These spacings, together with those in the ranges 12 Å–14 Å, 4.7 Å–5.2 Å, and 3.16 Å–3.4 Å, form a non-integral series, therefore, as Brindley has pointed out, a precise index cannot be assigned to any of these spacings. The 00 l spacings of ghassoulite and hectorite fall into this category also, as do many montmorillonites in an air-dried state.

Treatment of a montmorillonite with ethylene glycol usually causes expansion of the mineral in the c axis direction to give a basal spacing of ~ 17 Å. An x-ray diffraction pattern of such glycol treated material shows this spacing plus an integral series of higher orders from it. Figure 4 shows the patterns obtained from ethylene glycol treated oriented aggregates of hectorite, ghassoulite, and the two stevensite samples. Table 4 lists the spacings for first 6 orders of the 00 l series. No lines were visible beyond 006.

All samples depart to a greater or lesser degree from a strictly integral series. Brindley (1955, p. 245) has proposed that such departure in stevensite, together with the diffuse nature of the 00 l reflections, point toward the presence of nonexpanding material interstratified with montmorillonite type layers. He suggests that an interstratification of 9.5–10 Å and 14–15 Å structures could produce the irrational sequence of basal reflections and the 24–25 Å reflection observed in air dried specimens, and

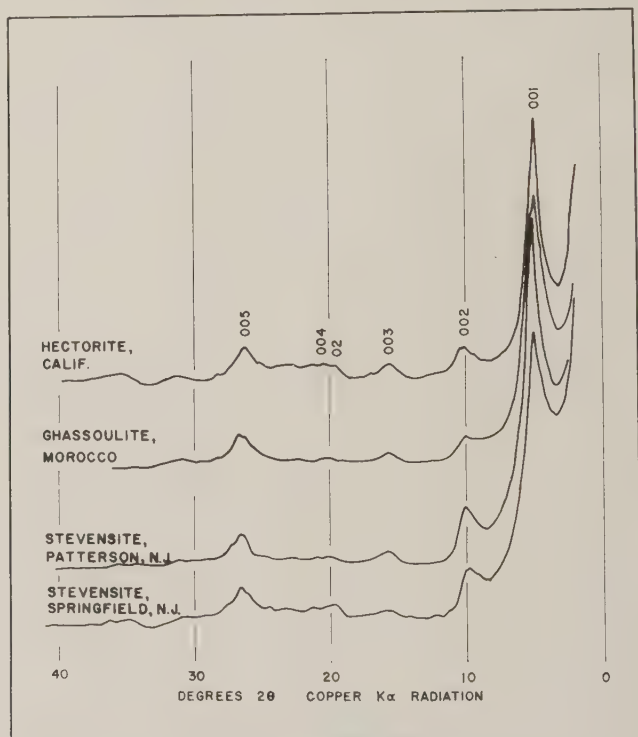


FIG. 4.—X-ray diffractometer patterns of oriented aggregates of hectorite, ghassoulite and stevensite treated with ethylene glycol.

that upon ethylene glycol treatment such a sequence of layers would give “a ‘first’ order at about 17 Å or a little higher spacing, a ‘second’ order at about 9 Å together with higher orders which cannot be reliably identified in the spectrometer traces.” If such is the case for stevensite then both hectorite and ghassoulite might be suspected of containing interstratifications of nonexpanding material. Neither shows a strictly integral sequence of basal orders, nor does either show a less diffuse or better developed sequence.

According to the diagram given by Weaver (1956 p. 206) an assem-

TABLE 4.—X-RAY DATA FOR ORIENTED AGGREGATES TREATED WITH ETHYLENE GLYCOL (Cu/Ni; $\lambda=1.5418 \text{ \AA}$)

Mineral	1. Hec- torite	2. Ghas- soulite	3. Steven- site No. 4	4. Steven- site No. 1	Calculated
Locality	Hector, Calif.	Morocco	Paterson, N. J.	Springfield, N. J.	
Indices	$d \text{ (\AA)}$	$d \text{ (\AA)}$	$d \text{ (\AA)}$	$d \text{ (\AA)}$	$d \text{ (\AA)}$
001	17.2	18.0	16.9	17.0*	17.0
002	8.8	8.8	8.8	9.0	8.5
003	5.69	5.68	5.69	5.68	5.66
004	?	?	?	?	4.25
005	3.40	3.37	3.39	3.37	3.40
006	2.86	2.90	2.87?	?	2.83

* The 15 \AA spacing for ethylene glycol saturated stevensite reported by Faust and Murata (1953, p. 976) was obtained only when the material was also saturated with "duco" cement thinned with acetone.

blage of mixed layered 10 \AA and 17 \AA units could contain up to about 30% expanded (17 \AA) layers without decreasing the first order spacing by more than 0.5 \AA . This would suggest that if any nonexpanding units are present in the materials under study, they probably account for less than 30% of any assemblage.

ELECTRON MICROSCOPE STUDIES

Electron micrographs of ghassoulite made at the University of Strasbourg show it appearing as laths and bands analogous to the laths and bands of hectorite. Electron micrographs of stevensite, on the other hand, resemble in morphology the common montmorillonites being irregular shaped thin plates and curled thin fragments (Brindley, 1955).

INFRARED ABSORPTION ANALYSIS

In order to test Brindley's (1955, p. 246) proposal that talc layers constitute the nonexpanding portions of stevensite, a method was sought that would differentiate between talc and other layer type structures that might also have a c axis spacing of 10 \sim 12 \AA . As talc seemingly alone among the layer lattice silicates gives an infrared absorption pattern with a strong absorption peak at a wavelength close to 15 microns, it was thought that any talc present in the samples studied might be revealed by the occurrence of this peak in the infrared spectra.

Robert G. Milkey of the U. S. Geological Survey kindly made infrared absorption analyses of talc, hectorite, ghassoulite, and Stevensite No. 4 as well as saponite, phlogopite, and pyrophyllite, close neighbors of talc

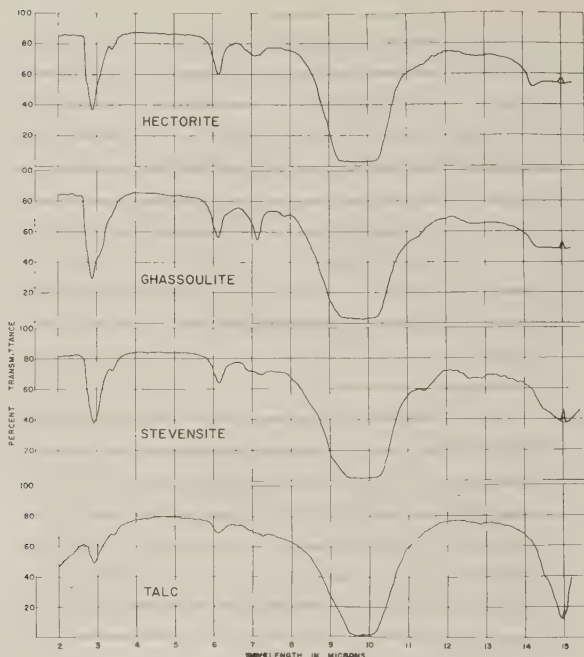


FIG. 5.—Infrared absorption spectra of hectorite, ghassoulite, stevensite and talc. Identical sample concentrations were used. The small sharp peak at 15 microns occurring in all the patterns is an instrumental effect. Robert G. Milkey, analyst.

in either composition or structure. The spectra obtained are shown in Figs. 5 and 6. Milkey's comments follow:

"Each of the minerals showed a strong absorption region at wave length 9 to 10 microns, which is shown by silicates in general. This peak, although showing some variations in shape among the different samples, was not distinctive enough to be useful for analysis of the structures.

"Other major absorption was due to hydroxyl and to the water molecules. These peaks are not necessarily quantitative, because no particular care was taken during the analysis to prevent absorption of water by the sample. It is of interest, however, that the hydroxyl peaks showed evidence of some non-hydrogen-bonded hydroxyl, together with stronger absorptions due to hydrogen-bonded hydroxyl which was largely accounted for as being present as H_2O molecules.

"The region of primary interest is at wave lengths from 14 to 15 microns. Talc has a strong absorption peak in this region at close to 15 microns, plus a weaker absorption peak near 14.5 microns. It seems likely that in this region, the absorption arises from the overall skeletal vibra-

tions of the crystal structure. Of the close neighbors of talc—stevensite, ghassoulite, saponite, hectorite, pyrophyllite, and phlogopite, only stevensite has a peak at this same point which is similar to that of talc. The intensity of the peak is somewhat less in the stevensite spectrum than in the talc. The other minerals showed only a diffuse absorption in this region: ghassoulite and hectorite showing a secondary peak near 14.4 microns, followed by a plateau of absorption, whereas saponite, pyrophyllite, and phlogopite had generally the diffuse plateau in this region.

"Spectra were also obtained of talc, and stevensite, ghassoulite, and hectorite which had been heated at 500° C. for 1 hour to eliminate inter-layer water and collapse the layers (Fig. 7). The heated sample of stevensite still showed, near 15 microns, the strong absorption peak, which now more closely resembled the talc peak than before. The ghassoulite and hectorite again did not show the strong peak at 15 microns; also, the minor peak which had previously appeared near 14.4 microns for ghassoulite and hectorite, was shifted after the layers were collapsed to a wave length of 14.5 microns. (It is interesting to note that this peak also occurs, as a weak 'shoulder,' in the spectra of stevensite and talc. In fact, the spectra of talc and stevensite in this region could possibly

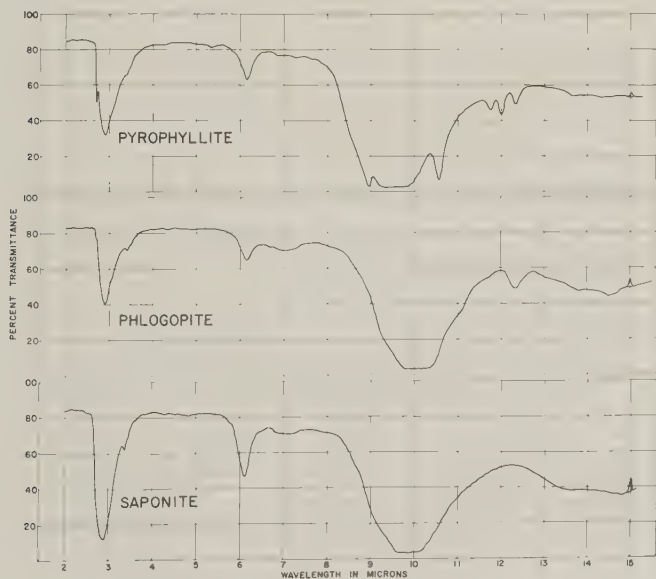


FIG. 6.—Infrared absorption spectra of pyrophyllite, phlogopite, and saponite. Identical sample concentrations were used. The small sharp peak at 15 microns occurring in all the patterns is an instrumental effect. Robert G. Milkey, analyst.

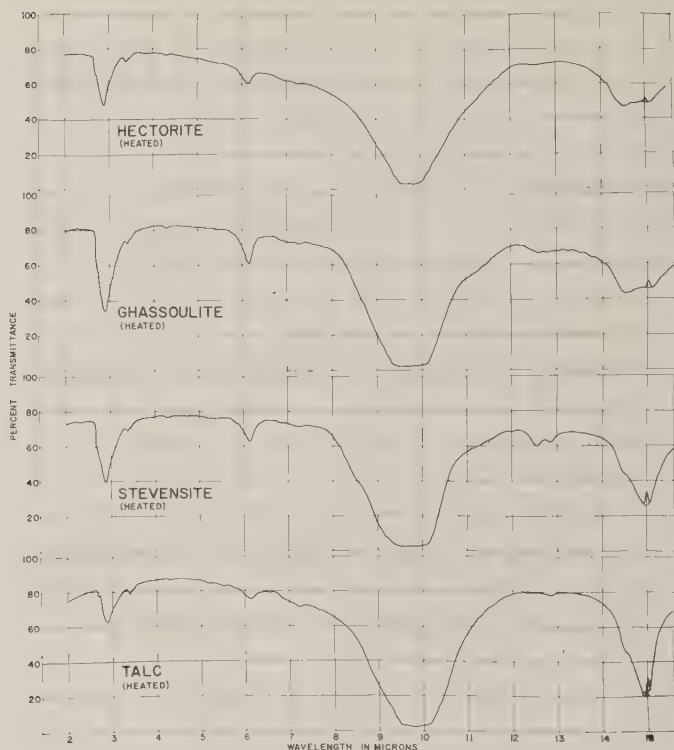


FIG. 7.—Infrared absorption spectra of hectorite, ghassoulite, stevensite and talc after each sample had been heated at 500° C. for one hour. Identical sample concentrations were used. The small peak at 15 microns occurring in all the patterns is an instrumental effect. Robert G. Milkey, analyst.

be the same as ghassoulite and hectorite, with the addition of the superimposed strong absorption at 15 microns.)”

On the basis of the infrared studies the data presented here do not unequivocally answer the questions regarding the presence of talc layers in the structures of stevensite. As Milkey states, “the particular type of bonding giving rise to the absorption in the 14.5 to 15 micron region is not known.” In addition the possible effect of the substitution of fluorine for hydroxyl in hectorite and ghassoulite has not been evaluated.

The significance of the infrared data for the heated and unheated specimens of talc and stevensite cannot be quantitatively assessed because the packing density was not determined and other quantitative factors were not controlled.

Bearing in mind these limitations, one could interpret the infrared data as suggesting that if non-expanding layers are present in stevensite they could be talc-like in nature.

SOLUTION STUDIES OF STEVENSITE, HECTORITE AND TALC

The ease of decomposing stevensite, and the other minerals now known to be identical with it, in hydrochloric acid has been known for a long time.

Samples of stevensite ground to the customary grain size for chemical analysis (-100 mesh) are completely destroyed by hydrochloric acid. Samples of talc similarly treated are not appreciably attacked. Murata¹ decomposed stevensite by leaching the mineral three times with 1:1 HCl. A sandy residue was formed. This residue, separated silica, is optically isotropic, has an index of refraction, n , of $1.45 \pm$ and is soluble in five per cent sodium carbonate solution.

Clarke and Schneider (1890) studied the solution of talc under the following experimental conditions—(1) heated in dry HCl gas to 383° – 412° C. for fifteen hours, (2) rapid evaporation with 75 cc. of fuming aqueous HCl, (3) eight days' digestion on the water bath with acid of 1.12 sp. gr., and (4) by digesting in like manner to (3) for thirty-two days. In the later experiment "3.94 (%) MgO and 0.41 (%) of sesquioxides were taken out." They conclude "Talc, therefore is, as should have been expected, remarkably stable in presence of hydrochloric acid, both aqueous and dry."

The marked contrast in the behavior of talc and stevensite under these experimental conditions is apparent.

Further data on the solution phenomena of stevensite, hectorite, and talc were obtained in this study.

The solution phenomena of stevensite were determined by digestion of the material in hot 10N HCl for about 6 hours. Two samples of stevensite were used and are referred to in the discussion as "1st run" and "2nd run." Both of these samples were prepared by gentle crushing, centrifuging and decantation. Mechanical deformation of the stevensite crystal plates has thus been kept at a minimum and the disorder in the crystal structure of the plates is not appreciably increased. Hectorite was likewise prepared and treated with acid. The results of these solution studies are recorded in Table 5. The column entitled "Wt. Loss in %, of original weight," in Table 5, includes not only the loss due to solution in the acid but also mechanical losses arising in filtering extremely fine-grained material which fails to settle or passes through the filter. This mechanical loss of sample gives rise to the discrepancy between the "MgO in %, of original wt." and the "MgO in typical sample of mineral."

The data for talc are not strictly comparable with those of stevensite for the samples were prepared by grinding dry for 30 hours in a mechanical mullite mortar. The starting material was a purified talc, U.S.P.,

¹ Murata, K. J., personal communication, 1953.

TABLE 5.—SOLUTION STUDIES OF STEVENSITE, HECTORITE, AND TALC IN HOT~10N HCl

Time of digestion in hours	Mineral	Wt. loss, in %, of original wt.	MgO in %, of wt. loss ¹	MgO in %, of original wt.	MgO in typical sample of mineral	Data on grain size	Notes
6	stevensite (1st run)	33.8	68.4	23.1	27.47	< 2 μ	stevensite No. 1 Springfield, N. J.
6	stevensite (2nd run)	32.5	62.5	20.3	27.47	< 2 μ	
6	hectorite	40.5	47.6	19.2	25.03	< 2 μ	Hector, Calif.
6	"ground talc" (no. T-1)			0.70	31.89	< 37 μ	Original sample
6	"ground talc" (no. T-1)			7.6	31.89	10-20 μ	Original sample ground 30 hours
6	"ground talc" (no. T-1)			10.5	31.89	2-10 μ	
6	"ground talc" (no. T-1)			11.1	31.89	1-2 μ	
6	"ground talc" (no. T-1)			16.1	31.89	< 1 μ	
60	"ground talc" (no. T-1)			0.62	31.89	< 37 μ	Original sample
60	"ground talc" (no. T-1)			8.5	31.89	10-20 μ	Original sample ground 30 hours
60	"ground talc" (no. T-1)			10.8	31.89	2-10 μ	
60	"ground talc" (no. T-1)			12.3	31.89	1- 2 μ	
60	"ground talc" (no. T-1)			17.0	31.89	< 1 μ	

¹ Harry C. Starkey, analyst.

Item 51-T-4350, from the Industrial Distributors which had already been ground to a fine grain size. A sample, no. T-1, ground for 30 hours give the following size analysis:

	Weight per cent
> 20 μ	1.6
10-20	7.8
2-10	38.0
1-2	10.5
< 1	36.3
	—
	94.2
Experimental loss	5.8

This size analysis is unusual in having two pronounced modes. Some of this unusual distribution of the grain size may be due to the presence of incompletely ground particles of talc which remained at the periphery of the mortar during much of the grinding process. The various data on the solution phenomena of ground talc are given in Table 5. Determinations of the amount of Mg in the solutions obtained from stevensite and talc were made by Harry C. Starkey and John C. Hathaway of the U. S. Geological Survey using the automatic photometric titration method of Shapiro and Brannock (1956, p. 41).

In discussing solution phenomena and comparing the results of the various experiments it is necessary to define the conditions of the studies. The data discussed here are solution studies of these minerals in hydrochloric acid and are not solubility measurements. From the standpoint of thermodynamics and the phase rule a given point on the solubility curve, at a given temperature, represents an equilibrium relationship involving the coexistence of the liquid phase and the solid phase. Such a relationship cannot be established in acid solution studies.

In such inequilibrium processes external characteristics such as grain size, perfection of the crystals or crystalline particles and other properties may greatly influence the rate of, or amount of solution.

Minerals of the montmorillonite group such as stevensite are generally composed of aggregates of tiny plates. When such an aggregate is placed in water or an aqueous solution it will break up and swell. In swelling the plates of stevensite will separate parallel to the basal plane and become smaller than they were in the original mass. In the processes of dispersion, centrifuging and decantation the original tiny plates have not only become thinner but their other dimensions may be reduced. In addition, the stevensite in water may undergo some ion-exchange with hydronium ions. That these tiny plates may come from larger plates is indicated by optical examination of stevensite which shows double refracting areas. These doubly refracting areas in stevensite represent optical continuity over areas sometimes as much as a few square millimeters.

Talc, on the other hand, has strong cohesive forces which prevent its mechanical disintegration by water and in order to reduce it to comparable grain size with that of stevensite it is necessary to grind it. The continuous dry grinding of this talc has probably introduced considerable disorder into the crystal structure of the individual plates of talc. Such disordered crystals would be expected to have a greater ease of decomposition than (1) plates slacked off larger plates and (2) synthetic crystals of the same size. Yoder and Eugster (1955, p. 256-257) have shown that for muscovite, grinding under almost identical conditions for thirty hours not only affects the particle size but also the shape and that measurable structural damage was observed.

Synthetic talc crystals grown as aggregates in hydrothermal synthesis studies by Bowen and Tuttle (1949) were examined by differential thermal analysis and *x*-rays and found to consist only of talc. This material was also examined with the electron microscope by Marie L. Lindberg, U. S. Geological Survey, who found the fully dispersed material to be uniformly thin tabular plates. The range in size of the fully dispersed particles is from about $.02\mu$ to 0.26μ . The size of the plates was judged by

reference to included spheres of polystyrene having a diameter of 2590 angstroms. Solution studies on this synthetic talc were made by Joseph J. Fahey. The sample was treated with 10N HCl for six hours on the steam bath. Magnesium was determined as the pyrophosphate. The synthetic talc yielded 15.7% MgO in the solution study which indicates that it was about 50 per cent decomposed by the acid.

Bearing in mind all of these limitations we may conclude that stevensite aggregates are decomposed by hot 10N HCl at all aggregate grain sizes from -100 mesh to the $<2\mu$ dispersed particle fraction. Hectorite is likewise destroyed. Talc on the other hand is more resistant, it is not mechanically disaggregated by water or aqueous solutions, and for particles from -100 mesh down to at least 37 microns it is relatively unaffected by hydrochloric acid. For samples which have been ground dry for thirty hours the decomposition of the "ground talc" increases with a decrease in grain size and for the material $<1\mu$ approximately one-half of it is decomposed under these conditions. These various experiments confirm the relative stability of talc over stevensite in the presence of hydrochloric acid.

Infrared spectra of the residue from the stevensite, 1st run, and the talc acid treatments are shown in Fig. 8. The talc diagram is essentially unchanged from that of untreated talc, but the stevensite is considerably altered. A pattern similar to that of amorphous silica occurs and the 15 micron "talc" absorption is absent.

CATION EXCHANGE CAPACITY

The total cation exchange capacity of samples of the stevensite No. 4 from Paterson, N.J., ghassoulite, and hectorite were determined by Dorothy Carroll of the U. S. Geological Survey using the colorimetric manganese method of Bower and Truog (1940, p. 411-413). The values obtained are given in Table 6.

The total cation exchange capacities reveal differences between the three minerals, ghassoulite and hectorite falling within the range usually associated with the montmorillonite group minerals, and stevensite falling below the range. This is in accord with the previous studies reported by Faust and Murata (1953). Talc has a cation exchange of 0.2 to 0.4 m.e. per 100 grams Kingery, Halden, and Kurkjian (1955).

The exchangeable ions in ghassoulite were determined by Harry C. Starkey and John C. Hathaway using the flame photometric and automatic photometric titration methods of Shapiro and Brannock (1956, p. 39-44). The results are given in Table 7.

These results show that exchangeable Mg plays an important role in ghassoulite and that the sum of octahedrally coordinated cations is probably less than 3.

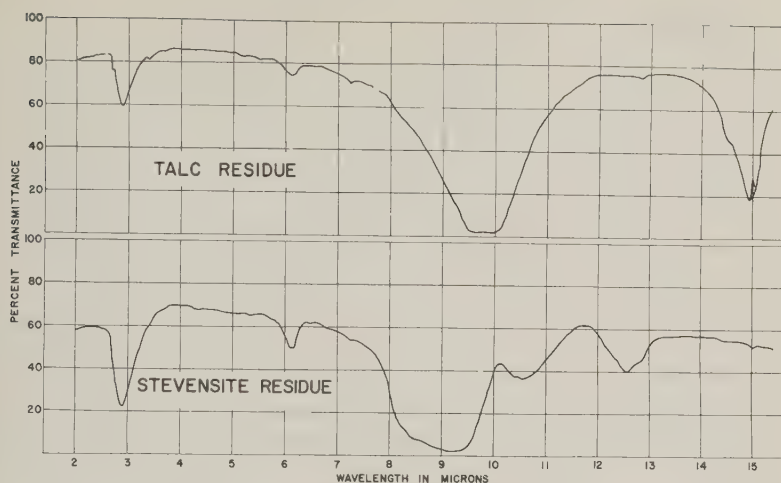


FIG. 8.—Infrared absorption spectra of residues from treatment of talc and stevensite with concentrated HCl. Identical sample concentrations were used. The small sharp peak at 15 microns occurring in all the patterns is an instrumental effect. Robert G. Milkey, analyst.

CHEMICAL ANALYSIS OF GHASSOULITE

Ghassoulite, analyzed by Augustin A. Damour in 1843 (Analysis 5, Table 1), contained 1.5% sand. The availability of ghassoulite from the studies of Millot (1954) afforded material for a new analysis. The new chemical analysis of ghassoulite, analysis No. 4 in Table 1, was made by our colleague, Joseph J. Fahey. Since the iron is chiefly in the ferrous state all of the iron is reported as FeO. The specific gravity of ghassoulite determined with a fused silica Adams-Johnston pycnometer using water as the filling liquid was found to be 2.336 at 4° C. The alkalies were determined by Joseph L. Dinnin using a flame photometer and the fluorine was determined by Sarah Berthold using a modified Willard-Winter method. It is of interest to note the general similarity between the

TABLE 6. TOTAL CATION EXCHANGE CAPACITY OF STEVENSITE, GHASSOULITE, AND HECTORITE

Mineral	Cation exchange capacity m.e./100 g.
Stevensite, No. 4, Paterson, N. J. <2 μ fraction	37.6
Ghassoulite, Morocco	75.1 ¹
Hectorite, California	60.8 ²

¹ Average of three determinations.

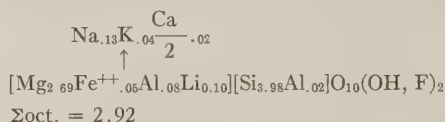
² Average of two determinations.

TABLE 7. CATION EXCHANGE RELATIONS OF GHASSOULITE

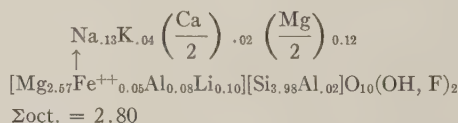
Exchangeable Cations	In m.e./100 g.	In per cent as oxides	
Na ⁺	17	Na ₂ O	.53
K ⁺	3	K ₂ O	.14
Ca ⁺⁺	2	CaO	.06
Mg ⁺⁺	55	MgO	1.13
Total	77		
Exchange capacity (from table 6)	75.1		

analyses of Damour and Fahey (note that 0.70% FeO \approx 0.78% Fe₂O₃).

To calculate the formula of ghassoulite 0.38% CaO was deducted to form with 0.30% CO₂, an amount of calcite equal to 0.68%. This gives 0.16% CaO as the amount in cation-exchange sites. Using the corrected analysis the formula is usually computed in the following form:



This formula must be modified however in the light of evidence given in Table 7 where Mg⁺⁺ is shown to be the principal cation-exchange ion. Using the measured cation-exchange for magnesium, one obtains the formula



The sum of 2.80 for the octahedrally coordinated ions is below that of either hectorite (2.98) or stevensite (2.86) and the charge on the structure is higher than in either of them.

INTERPRETATION OF THE DATA

In this study we have endeavoured to (1) find the relation between stevensite, ghassoulite, and hectorite and (2) to investigate further the interpretations of Brindley that stevensite is a talc-saponite interlayered mineral. All the methods we have used are not of the same sensitivity and for some techniques the results are not conclusive. It is desirable, therefore, to briefly summarize the results of each technique and to draw some conclusions from our research.

Differential thermal analysis studies

The DTA studies support the conclusions that stevensite, ghassoulite, and hectorite are all members of the montmorillonite group. The DTA curves for ghassoulite and hectorite up to 900° C. are more closely related to one another than to stevensite. Stevensite does not show a high temperature endothermic trough in the temperature range of 885° C. to 1061° C. where the dehydration of talc generally takes place. The absence of the high temperature endothermic trough, characteristic of talc, suggests that even if non-expanded layers are present in stevensite they are unlike talc.

Ghassoulite is also unlike talc in its behavior in DTA studies. Hectorite shows an endothermic trough beginning at 1053° C. extending as far as the record goes up to 1077° C. The significance of this trough has not been determined.

X-ray studies

The x-ray data for stevensite, ghassoulite, and hectorite reveal few differences between them other than the presence of a long spacing of about 25 Å in stevensite alone. As Brindley (1955, p. 242) has pointed out such a spacing could be explained on the basis of interstratification of 9.5–10 Å and 14–15 Å. Ethylene glycol treated oriented aggregates, however, gave patterns of essentially the same character for all of the minerals. The number and perfection of the basal reflections of hectorite and ghassoulite were little better than those of stevensite and all failed to show a strictly integral sequence of basal orders. Although the air-dried specimens of stevensite appeared to contain a large proportion of ~ 10 Å layers, less than 30% of non-expanded material would seem to be present in the glycol treated samples, otherwise a lower first order spacing would be expected.

The x-ray diffraction data thus give somewhat ambiguous results as to the differences between hectorite, ghassoulite, and stevensite.

Ghassoulite shows as little evidence of mixed layering as hectorite and like it exhibits no indication of "talc-like" layers upon collapse of the structure with heat treatment.

Infrared studies

Infrared absorption data on the untreated minerals show a distinct difference between stevensite and the other two minerals by the occurrence of the 15μ "talc" absorption. The increase in intensity of this peak on heat treatment indicates that if non-expanded portions are present in stevensite, this phenomenon could be associated with such portions but it

is also to be noted that the expanding portions are capable of producing this same absorption after they are collapsed by heat treatment. The loss of this absorption peak from the stevensite pattern but not the talc pattern after hot concentrated HCl treatment suggests that although the non-expanded portions exhibit some of the characteristics of talc, they are sufficiently different to be destroyed by acid treatment.

Chemical studies

Solution studies on talc, stevensite, and hectorite show the unstable nature of stevensite and hectorite in concentrated acid solutions, a characteristic of montmorillonite group minerals, whereas talc is quite resistant under these conditions.

Cation exchange studies on stevensite, ghassoulite, and hectorite show that ghassoulite and hectorite are most closely allied in that they exhibit a much higher cation exchange capacity than stevensite.

Chemical analyses of stevensite, ghassoulite, and hectorite show that ghassoulite and hectorite are more closely related to one another than to stevensite in that they both have significant amounts of fluorine and lithium. Moreover, they both contain some alumina.

We come now to the terminology of Brindley in referring to stevensite as a "talc-saponite interlayered mineral." The usage of saponite by Brindley to apply to a pure magnesium montmorillonite is not in agreement with established usage. Saponite was originally defined by Svanberg in 1840 as a hydrous silicate of magnesium and aluminum. All subsequent usage of the name has been in accord with this definition. Further, substitution of Al for Si in tetrahedral coordination is considered by Ross and Hendricks (1945, p. 47) to be a characteristic of saponite. The one closely allied mineral containing essentially no aluminum (0.13)% discussed by Ross and Hendricks was hectorite and although they group it with other Mg-rich trioctahedral montmorillonites they do not include hectorite under the term saponite. Inasmuch as the analyses of stevensite and synthetic stevensite show no alumina, by established usage, no part of stevensite can be considered saponite.

After the completion of this paper we received an offprint by Caillère and Henin (1956) on the subject of the nomenclature of saponite, aphrodite, rassoulite, and stevensite in which they conclude (translated from the French):

"It seems now advantageous to preserve two names to designate the extreme types of this mineral group: one corresponding to a large substitution, the other to a very small substitution in the tetrahedral layer. In this case, one may retain respectively the end members saponite and stevensite."

Genesis of ghassoulite

It is noteworthy that all of the specimens of stevensite and hectorite are of hydrothermal origin. On the contrary the "earth of ghassoul" is a sedimentary facies found in beds between the flint horizon and the marly limestone horizon. This authigenic mineral formed in a sedimentary basin under conditions analogous to those described by Millot, Radier, and Bonifas (1957). The great mineralogical similitude of hectorite and ghassoulite does not prevent their having different origins.

CONCLUSIONS

The results of the present research on stevensite and related minerals do not confirm Brindley's concept of stevensite as a talc-saponite inter-layered mineral and indicate that stevensite is a type of defect structure. From the cation exchange studies, Faust and Murata (1953) deduced that a charge deficiency apparently originates in stevensite through a deficiency in the total number of ions in octahedral coordination. Any portion of the stevensite structure which has such deficiency has the attributes of montmorillonite and any portion of the structure of stevensite which does not have a deficiency in the octahedral layers has the attributes of talc. Since the deficiency in the octahedral layers of stevensite is so great, stevensite is a montmorillonite group mineral. A random, or statistically disordered, distribution of the vacant sites in the sheets of octahedrally coordinated atoms would give a small proportion of domains with few or no vacancies and which have the attributes of talc. The size of such domains in stevensite is not known. Our experimental data indicate that such domains in the structure of stevensite which would have the attributes of talc occur in units too small to scatter x -rays coherently. Bradley (1950) has shown that the diffraction effects observed in this study can be expected to be produced by the type of assembly proposed above.

This concept of stevensite as a defect structure seems to fit all the experimental data as follows:

(1) The presence of domains with the attributes of talc in combinations with more abundant domains of stevensite could give rise to the weak reflection at $24.5\text{--}26.5 \text{ \AA} (\pm)$.

(2) The appearance of the infrared absorption peak near 14.5 microns in natural stevensite and its disappearance in the residue of acid treated stevensite.

(3) The absence of an endothermic trough corresponding to talc in the stevensite DTA pattern.

(4) The complete decomposition of stevensite in acid.

The alternative explanation of Brindley that stevensite is made up of discrete interlayers of talc in saponite could be used to explain the occurrence of a weak reflection at 24.5–26.5 Å but it fails to explain observations 2, 3, and 4.

Theoretically one could derive the defect structure of the trioctahedral montmorillonite stevensite by creating, in random distribution, the maximum number of vacancies in the octahedral sheet (brucite sheet) in the talc structure and still preserve a stable crystal structure. Analogously, one could derive the defect structure of the dioctahedral montmorillonite by creating in random distribution the maximum number of vacancies in the octahedral sheet (gibbsite sheet) in the pyrophyllite structure. Rustum Roy and L. B. Sand (1956) have claimed the synthesis of this "defect pyrophyllite" structure which they called pure aluminum montmorillonite.

The defect structure theory proposed here is another way of deducing the dioctahedral and trioctahedral montmorillonite structures using as prototypes-talc and pyrophyllite. The original usage of talc and pyrophyllite as prototypes for montmorillonites was proposed by Hofmann, Endell and Wilm (1933) and modified by Marshall (1935) and Hendricks (1942). This theory deduces the structure of montmorillonite from the structure of pyrophyllite and talc through the replacement of silicon in tetrahedral coordination by trivalent aluminum, or by the replacement of aluminum in pyrophyllite or magnesium in talc by other ions, or by a combination of both types of replacements.

Vacant sites in the trioctahedral montmorillonite-sauconite have been discussed by Faust and Murata (1953).

RELATIONSHIPS AMONGST STEVENSITE AND SOME SIMILAR MINERALS OR MIXTURES

In addition to the minerals studied by us, other investigators have reported on related minerals, or on the synthesis of hydrous magnesium silicates which are discussed below in the light of our own studies.

Syntheses of stevensite

Strese and Hofmann (1941) were probably the first to synthesize stevensite. Their preparation "Synthetischer Magnesium montmorillonit 33B" gives an *x*-ray pattern comparable with that of natural hectorite. Their experimental data suggest that for most of their preparations K^+ made up most if not all of the cation exchangeable ions; in some it was Ca^{++} and probably also Mg^{++} .

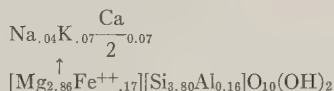
Bowen and Tuttle (1949) in their study of the system $MgO-SiO_2-H_2O$ attempted to synthesize the mineral sepiolite. They were apparently not

entirely satisfied with the identification. They based their preparation on the formula $2\text{MgO} \cdot 3\text{SiO}_2 \cdot 2\text{H}_2\text{O}$ which differs from stevensite in having about 3% more silica and about 3% less water. They mention (p. 441) that the experimental charges lost silica by transport in the vapor phase and accordingly the formation of stevensite in their experiment is not unexpected. X-ray and DTA studies of a large sample prepared by them for Faust and reported on in this paper shows the charge to consist of stevensite with a very small amount of chrysotile.

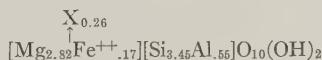
Roy and Roy (1955) accomplished the synthesis of stevensite in the system $\text{MgO-SiO}_2\text{-H}_2\text{O}$ and Mumpton and Roy (1956) studied the stability temperature of this synthetic stevensite.

Interstratified "saponite-talc" of Alietti

Alietti (1956) described an interlayered saponite-talc mineral from Monte Chiaro, Italy and his chemical analysis is given in Table 1. The original paper contained a misprint for the value of silica of 5.31, which was corrected in ink to 54.31. A recalculation of this analysis yielded the formula



which does not agree with that given by Alietti namely



There may be some error in the reporting of the data.

On the basis of our calculations there is no deficiency in the octahedral layer ($\Sigma=3.03$) and this mineral could have no stevensite layers in its structure. However, as we showed in the section on the formula of ghasoulite, cation exchangeable magnesium must be determined on minerals of this kind in order to determine the actual composition of the octahedral layer. Until this is determined one cannot use Alietti's chemical analysis to establish the presence and proportions of stevensite-type or saponite-type layers in the structure. An examination of Alietti's other data does confirm his interpretations that the mineral he studies is a montmorillonite-type with regularly interstratified non-expanding layers, in the proportion of 1:1 as judged by his diffractometer patterns, 1b and 1c (p. 203).

Hanušite

Kašpar (1942) studied a hydrous magnesium silicate which he called hanušite. Hanušite occurs as yellowish-white to yellow-brown radiating aggregates, similar to pectolite, in the tholeiite of the Liebstadt region

in the Riesengebirge. Kašpar concluded that hanušite was the magnesium end member in the supposed series pectolite-walkerite. Kašpar was unaware of the work by Glenn (1916), or he would have recognized that stevensite had already been assigned to the name of the magnesium end member of this supposed series. An examination of Kašpar's data leaves little doubt that hanušite is a mixture of stevensite and pectolite. Thus, there is no alumina in hanušite, it is completely decomposed by HCl, the dehydration curve is similar to that of montmorillonites and the chemical composition after subtraction of hypothetical pectolite from the analysis yields a composition of the right order of magnitude for stevensite. Using the CaO value in Kašpar's analysis and Fig. 1 in Faust and Murata (1953) hanušite appears to be made up of about 88% stevensite and 12% pectolite.

RECOMMENDATION

It is recommended that the names hanušite (=stevensite+pectolite) and ghassoulite (=hectorite) be relegated to the synonymy.

ACKNOWLEDGMENTS

We are indebted to Drs. Paul F. Kerr, Clarence S. Ross, O. Frank Tuttle, and S. C. Williams for specimens used in this study. The infrared studies were made by Robert G. Milkey, U. S. Geological Survey. The various chemical analyses and determinations were made by Joseph J. Fahey, Charles Milton, Dorothy Carroll, Harry C. Starkey, Joseph I. Dinnin, and Sarah Berthold, all of the U. S. Geological Survey, to whom we express our sincere thanks. The electron micrographs of synthetic talc were prepared by Marie L. Lindberg, U. S. Geological Survey.

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MEMORIAL OF CARL FAESSLER

P. E. AUGER, *Université Laval, Québec, Canada.*

With the death of Carl Faessler in Quebec City on October 1st, 1957, the American Mineralogical Society lost a most distinguished member, at the same time the Faculty of Science of Laval University was losing the dean of its professors and Canada, a great citizen.

Born on the 24th of September 1895 at Steinen in Switzerland, Carl Faessler received his fundamental education in the schools of Feusisberg and Steineberg. He obtained his Bachelor of Arts at the College Maria Hilf of Schwyz in 1916. In 1917 he registered at the University of Fribourg where he was especially interested in Chemistry, Botany and Min-



CARL FAESSLER

1895-1957

eralogy and where he obtained in 1923 a Doctorate of Science *Summa Cum Laude*.

It was in 1923 that Dr. Faessler came to Canada to act as assistant professor of Chemistry and to take charge of the courses in Mineralogy and Geology at the newly founded *Ecole de Chimie* of Laval University in Quebec.

In 1931, Dr. Faessler was appointed full professor, responsible for the teaching of Mineralogy, Crystallography and Petrography. He also gave courses in General Geology, Mineral Technology, and even in German.

His capacity for working was tremendous. He had a very systematized mind and every minute available was used for something useful.

Besides his teaching assignment Faessler spent a good part of his time and energy doing field work for the Department of Mines of the Province of Quebec. He was one of the first field geologists for the Province and his association with the Provincial Government extends from 1928 to 1945. All those who worked with him in the field recall his untiring energy and his devotion to his work. His sense of discipline and order was carried to the most minute details, and in the most difficult situation one could not help feeling a sense of security radiating from his strong and powerful personality.

He was a leader and yet he was first and above all a teacher and professor. Everyone that was associated with him at the University, in the field or in a social gathering was fascinated by the tremendous knowledge that he had acquired through abundant reading, personal experience and profound philosophical thinking.

Dr. Carl Faessler during his 34 years of active life as a geologist has taken part in the activities of a great number of Scientific Societies.

All his publications cannot be listed here due to lack of space, but his last work should be mentioned, *The Cross Index to the Geological illustrations of the Geological Survey and the Mines Branch*; the Ontario Department of Mines and the Quebec Department of Mines. These are extensive pieces of work which took years to complete and which are considered by everybody as a great contribution to Canadian Geology.

MEMORIAL OF FRANK FITCH GROUT

GEORGE M. SCHWARTZ, *University of Minnesota, Minneapolis, Minn.*

Frank Fitch Grout died in Minneapolis August 1, 1958. He had not been well for over a year, but worked in his office at the University almost daily until he collapsed on July 26th.

He was born in Rockford, Illinois January 24, 1880, the son of Charles Leroy and Louise Fitch Grout both originally residents of New Hampshire. He graduated from the high school course at Throop Polytechnic Institute, the forerunner of California Institute of Technology. He also started college work at Throop but finished his undergraduate work at the University of Minnesota where he received his B.S. degree in 1904 with a major in chemistry. He changed to geology and received the Master of Science degree in 1908 and the degree of Doctor of Philosophy from Yale in 1917. Grout served briefly on the geological surveys of West Virginia and Illinois and as an instructor at the University of Oklahoma. He joined the staff of the geology department at the University of Minnesota in 1907 and held appointments continuously until his retirement as Professor Emeritus in 1948. His rise in rank was rapid and he was promoted to Professor in 1919. In 1906 he married May W. Browne who survives him. He is also survived by a daughter Jean Grout Condit and four grandchildren, and by a sister Mrs. Myra Christie.

Frank Grout's main interest and research was in the field of petrography and petrology but his interests were broad and his contributions included papers on coal, clays, iron formations and ores, mineralogy, chemical analysis of rocks and Precambrian stratigraphy. Undoubtedly his greatest love in geology was for field work in the lake country of northern Minnesota and adjacent areas in Ontario, where he could travel by canoe and portage, camping in the less accessible places where outboard motors usually failed to trespass. Here he enjoyed nature in the broad sense as well as geology, and to work with him was a privilege never to be forgotten. It is fitting that his ashes were spread on the waters of Saganaga Lake.

Part of Grout's great success in his professional life was due to the fortunate association with William H. Emmons. Professor Emmons came to the University of Minnesota in 1911 as Chairman of the Department of Geology and Director of the Minnesota Geological Survey and immediately recognized Grout's outstanding ability. Emmons was busy on various consulting problems during the summer and delegated supervision of summer field work for the Minnesota Geological Survey to Grout. For a period of 35 years Grout worked on Minnesota's geology and supervised the work of younger staff members of the Department of



FRANK FITCH GROUT

1880-1958

Geology and graduate students. The large output of reports which resulted from the very small budget of the Minnesota Survey was, in no small part, due to Grout's ability and hard work. In the field his enthusiasm knew no bounds and he would return time and again to an important outcrop to try to pry out its secrets. He was a quiet but masterful teacher both in the classroom and in the field. Dozens of young geologists learned thorough methods of field work under his guidance.

His reputation as a teacher is shown by the fact that in successive years after retirement he taught at the Universities of Florida State, Columbia, Arizona, and California Institute of Technology.

In both field work and research Grout was a philosopher rather than a technician. He was interested in methods only to the extent necessary to obtain needed data. His active mind was concerned with the meaning of observations and their bearing on the origin of rocks and minerals. The writer worked with him longer and more closely than anyone and never ceased to be amazed at his penetrating insight into the problems of complex Precambrian geology. To have been his student, coworker and friend was a great privilege.

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MEMORIAL OF ELBRIDGE CHURCHILL JACOBS

ROBERT K. DOTEN, *The University of Vermont, Burlington, Vermont*

Elbridge Churchill Jacobs, former State Geologist of Vermont and a member of the faculty of the University of Vermont for nearly fifty years, died of a cerebral hemorrhage at Burlington, Vermont on December 12, 1957 at the age of 84.

He was born at Ogunquit, Maine on February 15, 1873. He attended the Portsmouth, New Hampshire high school and went on to get his B.S.



ELBRIDGE CHURCHILL JACOBS

1873-1957

degree from the Massachusetts Institute of Technology in 1896 and his M.A. degree from Columbia University in 1897. For the next two years he served as an assistant in the department of Mining and Metallurgy at M.I.T. and was then appointed an instructor in Chemistry and Mineralogy at the University of Vermont in 1899. He was advanced to assistant professor in 1901 and to full professor in 1904. In 1924 he was instrumental in establishing a department of Geology at the University and was appointed Professor of Geology and Mineralogy to head up this newly formed department. He became Professor emeritus in 1944, but continued to teach geography to the United States Air Force Detachment at the university until the end of World War II.

He was appointed State Geologist for Vermont in 1933 and retained this position until he retired from it in 1946. During this period he published six biennial reports to which he was the principal contributor.

The year 1933 was also noteworthy in that it marked the beginning of observations at the University of Vermont Seismograph Station. This station was established through cooperation of the University of Vermont and the National Research Council and came into being very largely through the efforts of Professor Jacobs. He was the director of the station from its establishment until it was dismantled in 1957. He also became curator of the mineral and geologic collections of the university in 1933.

During the flood control work undertaken by the U. S. Corps of Engineers in western Vermont in 1929 following the disastrous floods of 1927, he served as consulting geologist.

Professor Jacobs' published works deal largely with the geology and mineral resources of Vermont. His first paper was a study of the talc deposits of Vermont published in the ninth Report of the Vermont State Geologist in 1914 and his last, a book entitled "The Physical Features of Vermont" published by the Vermont State Development Commission in 1950. He was particularly interested in the chemical aspects of geology, and a number of his papers carry original rock and mineral analyses made by him. In his later years, he did much field work in connection with the structure of the northern part of the Green Mountains; only part of which he was able to complete and have published.

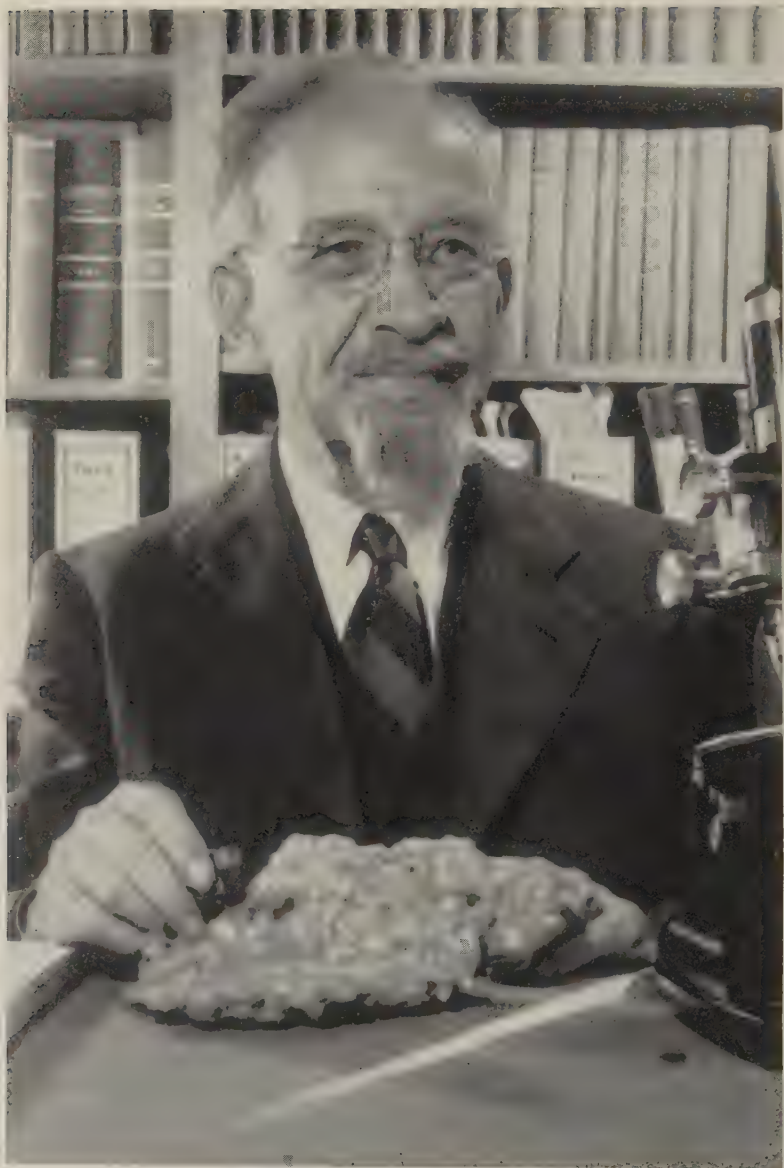
He was a fellow of the American Association for the Advancement of Science, the Geological Society of America, and the Mineralogical Society of America; a member of the American Association of University Professors, the American Geophysical Union, the Polar Society, and the Seismological Society of America.

He is survived by his wife, Mrs. Jessie Louise (Chapman) Jacobs; a

daughter, Mrs. G. H. Klinck; and two grandsons, John and Daniel Klinck of Washington, D. C.

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ALEXANDER NEWTON WINCHELL

1874-1958

MEMORIAL OF ALEXANDER NEWTON WINCHELL

R. C. EMMONS, *University of Wisconsin, Madison, Wisconsin.*

Alexander Newton Winchell, Emeritus Professor of geology, died in New Haven, Connecticut on Saturday, June 7, after a short illness. He was 84 years old. He was born in Minneapolis, Minnesota, on March 2, 1874, the son of a distinguished geologist, N. H. Winchell. He received his B.S. degree in 1896 and his M.S. degree in 1897 from the University of Minnesota, and his D.Sc. from the University of Paris in 1900.

Professor Winchell served as Professor of Mineralogy and Petrology at the Montana School of Mines from 1900 to 1907, and as Professor of Geology at the University of Wisconsin from 1907 till his retirement in 1944. He was chairman of the Department of Geology from 1935 to 1940. He was associated with the U. S. Geological Survey from 1901 to 1910. After his retirement he served as Visiting Professor of Mineralogy at the University of Virginia 1948-49 and at Columbia University 1949-50. He served for three years as resident mineral consultant at the Stamford laboratory of the American Cyanamid Company.

Professor Winchell was a fellow of the Geological Society of America, the Mineralogical Society of America, the Society of Economic Geologists, the American Association for the Advancement of Science and the Wisconsin Academy of Science. He was councillor of the Mineralogical Society of America from 1927 to 1930 and was president in 1932. He was awarded the Roebling medal of the Society in 1955 for meritorious achievement.

Professor Winchell was best known for his books on optical mineralogy which have served the science eminently for many years. The first edition was written jointly by him and his father, and the last, fourth, edition was written in collaboration with his son, Horace. These books appeared in three volumes,—volume 1 being devoted to optical crystallographic theory, volume 2 a very thorough compilation of published data on the optical properties of mineral materials including many charts of his own design on mineral properties, and volume 3, a set of exhaustive tables for the identification of minerals by their properties. His volumes 2 and 3 are unexcelled in any language in their thoroughness and value to the laboratory worker. They appeared also in a Russian edition of 15,000 copies. Immediately prior to his death, he visited Moscow, as is required by Russia, in relation to the royalties on the Russian edition.

In addition to his textbooks on optical mineralogy Professor Winchell compiled a text on the optical properties of artificial minerals and one on organic crystals. He also compiled a textbook on mineralogy.

The major research activity of Professor Winchell is embodied in a

considerable list of published papers on the correlation of the optical properties of crystalline materials and their chemical composition. The graphical representation of these findings will long be a serviceable monument to his devotion to these studies.

During his long teaching career, Professor Winchell made many friends among students who will recall the thoroughness of his training and who recognize the role that his influence has played in their professional accomplishment. His absences from the national meetings were rare and inevitably evoked inquiries on his welfare.

Professor Winchell is survived by his wife, Florence S. Winchell, his two sisters, Mrs. A. W. Grant of St. Charles, Ill., and Mrs. B. J. Denman of Minneapolis, Minn., and by four of his five children—Mrs. R. E. Rettger of Dallas, Texas., Dr. A. V. Winchell of Rochester, N. Y., Mrs. R. J. Lund of Columbus, Ohio, and Professor Horace Winchell of New Haven, Conn. A memorial service was held at Spring Glen Congregational Church in Hamden on Tuesday, June 10.

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- In addition to these there are several unpublished mining reports. He also cooperated with N. H. Winchell in the publication of a genealogy of the Winchell family.

MEMORIAL OF SAMUEL ZERFOSS

A. VAN VALKENBURG, *National Bureau of Standards, Washington, D. C.*

Dr. Samuel Zerfoss, a Fellow of the Mineralogical Society of America and Chief of the National Bureau of Standards Refractories Section, died on December 19 from injuries received in a tragic pre-holiday automobile accident in Washington, D. C. The day before he and two colleagues were enroute to Baltimore where Dr. Zerfoss was to have been honored with the office of Chairman Elect of the Baltimore-Washington section of the American Ceramic Society. Moments after leaving the Bureau of Standards an irresponsible driver raced through a stop sign and crashed into the official car containing Dr. Zerfoss and his associates.

Dr. Zerfoss was born on November 1, 1912, at Hummelstown, a small community located in the hills of Pennsylvania. The rocks, the minerals and the ores of the nearby hills undoubtedly did much to influence his interests in the earth sciences. In 1930 he began his formal education at the School of Chemistry and Mineral Industries at the Pennsylvania State College. Here he received a Bachelor's degree in chemistry in 1934 and a Master's degree in 1936. His potential as a research worker and as a teacher was early recognized by the university and he was invited to become a member of the staff. From 1937 to 1947 he held the Bethlehem Steel Fellowship doing research on refractories, slags, high temperature mineralogy, and chemistry. A Doctor's degree in ceramics was awarded him in 1941. Dr. Zerfoss was made an Associate Professor of Ceramics in 1942, teaching both graduate and undergraduate courses. In addition to his work with students, this was a period highly productive in personal research. His contributions ranged over the areas of microscopy, phase equilibria, chemical analysis, mineralogy, and metallurgical inversions. In addition to six major publications, he supervised an industrial fellowship and was involved in numerous problems concerned with stimulating industries utilizing Pennsylvania natural resources.

In 1947 Dr. Zerfoss accepted a position with the Naval Research Laboratory in Washington, D. C., to head a program of fundamental research on a wide variety of materials preparation including crystals, glasses, and ceramics needed for both defense and commercial peacetime uses. It was during this period that he established for himself, his colleagues, and the laboratory an international reputation for materials research, especially in the field of crystal growth. His publications on the fundamentals of crystal growth are classic to this day.

In April 1955 Dr. Zerfoss joined the staff of the Mineral Products Division of the National Bureau of Standards. Four months later he was appointed Chief of the Refractories Section, where he directed a program



DR. SAMUEL ZERFOSS

1912-1958

of fundamental research on the properties of materials at high temperatures. His reputation in this field grew with each passing year, and in 1956 he was appointed a U. S. member of the International Union of Pure and Applied Chemistry, and a member of the High Temperature Committee of this group. Dr. Zerfoss contributed his time and his talent to several scientific organizations. He was a Fellow of The American Ceramic Society and The Mineralogical Society of America. He held membership in The American Chemical Society, The Geological Society of Washington, The Washington Academy of Science, The Chemical Society of Washington, The Geochemical Society, The Societe Chimique de France, Sigma Xi, and Sigma Pi Sigma.

In personal life Dr. Zerfoss was known to all his friends as Sam. He was modest in his approach to others, humble in his evaluation of himself, and generous to those who needed advice or help. Sam, a bachelor, was an avid collector of books, magazines, records, minerals and, most important to him, friends. He took great delight in entertaining at his new home that he recently bought. A dinner, complete with all the trimmings, prepared and served by Sam was a topic for continued conversation. He was a member of several groups outside the scientific community; to mention a few, The Washington Opera Society, The Museum of Modern Art, The Theater Guild, and The Metropolitan Opera Guild.

Dr. Zerfoss is survived by a sister, Martha, of Philadelphia, and two brothers, George Zerfoss, a Nevada mining engineer, and Commander Allen Zerfoss, U.S.N., San Pedro, California.

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MARTIN J. BUERGER

PRESENTATION OF THE ROEBLING MEDAL TO MARTIN J. BUERGER

CLIFFORD FRONDEL, *Harvard University, Cambridge, Massachusetts.*

Mr. President, Fellows and Members of The Mineralogical Society of America, and Guests: It is a pleasure to introduce Martin J. Buerger as Roebling Medalist. This award, the highest honor that our Society can bestow, is now given for the first time to an American structural crystallographer.

The development of structural crystallography through x-ray and other techniques has brought fundamental advances in our understanding of minerals and in our methods of dealing with them. W. L. Bragg, a pioneer in structure analysis, received the Roebling Medal in 1948. Paul Niggli, who worked in cognate fields, as they were relevant to his main interests in petrology and mineralogy, was awarded the Roebling Medal in 1947.

Buerger has worked broadly in this field. His main interest has been the application of structural data and of crystallochemical principles derived from structural studies to the interpretation of genetic problems in mineralogy and geology. Yet interpretations stem from experimental observations, and are tested by them. It is from this realization that the great bulk of his work has been concerned with instrumentation, particularly the Weissenberg and precession methods of single-crystal goniometry and powder photography by film methods, with crystallographic theory and the methods whereby the structure of a crystal is derived from the observed diffraction effects, and with the description by these means of the crystal structure of numerous substances.

Buerger's penetrating interests in the thermodynamics and structural control of polymorphism, in disorder and exsolution, and in thermostructural phenomena in petrology and geochemistry are indicated in the notable Presidential Address that he gave at the Ottawa meeting of our Society in 1947. In this Address I recall the thematic statement "Temperature endows a mineral with energy beyond that of its static crystal structure." A forthright and clearspoken person, not averse to rigor in any of its forms, Buerger sometimes has endowed people with temperature. His past laboratory assistants include charred examples—yet without exception their feelings became transmuted with time to affection.

Exacting thoroughness in preparation and demonstration, often combined with originality of method, is a characteristic of the man. His book "X-Ray Crystallography" treating chiefly of the Weissenberg method and his recent "Elementary Crystallography" are typical examples.

M. J. Buerger was born in Detroit in 1903. His undergraduate and

graduate college work was done at the Massachusetts Institute of Technology. He was appointed Assistant Professor at Tech in 1929, and became Professor of Mineralogy and Crystallography in 1944. Appointed Institute Professor in 1956, he is now Director of the School for Advanced Study.

A Fellow of the American Academy of Arts and Sciences, and a member of the National Academy of Science, he is a member of numerous other scientific groups including the Academies of Science of Torino and of Brazil. He founded the Crystallographic Society of America, that later merged with the American Society for X-Ray and Electron Diffraction, and was President of the latter organization in 1939. He has served the International Union of Crystallography both as a Council member and as a member of the U. S. National Committee. Our delegate to the organization meeting of the International Mineralogical Association, in Madrid last April, he performed his duties in exemplary fashion, participated in Spanish cultural activities and, in the always welcome discussion of the distribution of offices between countries, poured cement on Anglo-American relations. He is a co-editor of the *Zeitschrift für Kristallographie*, and a co-editor of the *International Tables for X-Ray Crystallography*. In 1951 Professor Buerger was awarded the Day Medal of the Geological Society of America for the "distinguished application of physics and chemistry to geology." He has long served our Society; he has given direction and impetus to Mineralogical Science. Mr. President, it is an honor to present Martin Julian Buerger for the award of the Roebling Medal of the Mineralogical Society of America.

ACCEPTANCE OF THE ROEBLING MEDAL OF THE MINERALOGICAL SOCIETY OF AMERICA

MARTIN J. BUERGER, *Massachusetts Institute of Technology,
Cambridge, Massachusetts.*

Mr. President, Professor Frondel, Fellows, Members and Guests of The Mineralogical Society of America:

The award of the Roebling medal is a great honor, and I accept it with great pride, for this is my own professional society. There are few satisfactions in life which compare with recognition in one's chosen field. At the same time, I wish to express my feeling of humility, for gathered in this room are many mineralogists who put my own knowledge of mineralogy to shame.

I thought that you might be interested in a sketch of the changing complexion of mineralogy as I have observed it. I became a graduate student in geology in 1925, and attended my first meeting of this society when it met in Cambridge in 1932. As I joined this group we argued about the relative importance of crystal habit and internal structure. We continued the discussion at later meetings. It was a sign of the times. Mineralogy was just beginning to change from a descriptive science to the first science to recognize what is now popularly called *the solid state*. From about that period, detailed descriptions of crystal habit began to give way to another kind of description—that of internal arrangement of atoms. This in itself was simply a change from one descriptive science to another, brought about by a change of tool. But the new description of minerals brought with it the power to explain their properties and occurrence, and soon there was a clearer understanding of such things as crystal growth and habit, twinning, plastic deformation, phase transformations, reaction series, and other features of minerals which had been known but not really woven into a science. I was fortunate enough to be one of the youngsters who were on the scene as this development took place, and it was fun taking part in it. There is still a lot to be done, and still a lot of fun to be had.

The importance of crystal structure not only in mineralogy but in chemistry, physics, and metallurgy, made the working out of the arrangements of atoms in crystals a very important matter during this period. This was something of an art, until what is now known as the *phase problem of x-ray crystallography* was appreciated and formulated. Dozens of us became involved in attempts to solve this problem. While a true general solution of the problem has not been achieved, the general nature of its restrictions is now pretty clear, and it can be said that most, if not all, mineral crystal structures can be solved within its restrictions. It is

surprising how many mineral structures are now known, and certainly the science of mineralogy is in a position to advance by taking advantage of the known structures of minerals. I am sure our next great advances will come along theoretical lines, and will include, among other things, a sufficient understanding of the relations of phase diagrams to the structures involved such that at least some phase diagrams can be predicted. We are also on the brink of studying the mineralogy of high pressures, that is, the mineralogy of the deep crust. In another field, we can expect a more thorough understanding of the significance of crystal habit in mineral paragenesis. I hope that I may still participate in the development of some of these inviting subjects.

Returning to the Roebling medal which you have so kindly bestowed upon me, I would like to reiterate how much I value this honor, for it signifies that some of the things which I thought interesting enough to study and to write about, you also thought interesting enough to read about. To a scientist there is no more sincere compliment than to indicate that there is something interesting in what he has done. I deeply appreciate this compliment and will always value this splendid token of it.



CHARLES EDWARD WEAVER

PRESENTATION OF THE 1958 MINERALOGICAL SOCIETY OF AMERICA AWARD TO CHARLES EDWARD WEAVER

RICHARDS A. ROWLAND, *Shell Development Company, Houston, Texas.*

Mr. President, Fellows, and Members of The Mineralogical Society of America, and Guests: It is a double pleasure to have the privilege of presenting to you my friend and colleague, Dr. Charles Edward Weaver, as the recipient of the seventh annual award of the Mineralogical Society of America, because he is a clay mineralogist and is associated with an industrial research laboratory. In the past decade there has been a rapid awakening throughout the industry, and especially the petroleum industry, to the valuable contribution which mineralogists are able to make in industrial research. To the individual mineralogist undertaking a career in industrial research the problem has been twofold; first to solve a scientific problem and second to sell the solution by showing its actual application and usefulness. It is therefore, most gratifying that the recipient of this year's award has been able to fulfill both of these requirements and in so doing to make scientific contributions worthy of recognition by our Society.

A graduate of the Pennsylvania State University, whose young Ph. D's are as well known for their aggressive attitude as their superb scientific training, Chuck Weaver joined the Exploration and Production Research laboratory of the Shell Development Company during a period of expansion to work on a specific operational problem. Until then clay mineralogists spoke of kaolinite, illite and montmorillonite. Soon we all knew about K-bentonite, 2-1, 4-1, 7-3, and other mixed layers. These are pretty good odds.

Because so little was known about the distribution of clay-minerals in sedimentary basins, an immense data gathering program was first necessary. This involved the use of x-ray diffraction and other determinative techniques on more than 20,000 samples. With his quick perception, hard work, and crusader's spirit, it was not long before the clay mineral literature in the form of company reports had more than quadrupled. Stock rose fifty per cent and oil reserves reached an all-time high. I am not certain that Chuck was responsible for all of this, but I know that for a while he wrote about every third company report. And this while, by his own admission, he was learning to write the English language, an art for the teaching of which there is not time at his alma mater. Because of an enlightened and sympathetic management, much of the best and most fundamental of Dr. Weaver's work was soon made available for publication in scientific journals.

President Goodspeed, it is a pleasure to present Dr. Charles Edward Weaver, as the recipient of the seventh annual award of the Mineralogical Society of America.

ACCEPTANCE OF THE 1958 MINERALOGICAL SOCIETY OF AMERICA AWARD

CHARLES E. WEAVER, *Shell Development Company, Houston, Texas.*

Mr. President, Fellows, and Members of the Mineralogical Society of America and Guests: As a scientist in industry I am especially grateful to the Society for extending this recognition to me, for in doing so you are recognizing the increasingly important contributions being made by mineralogists associated with industrial research organizations. Such contributions are possible because many large companies such as Shell Oil Company have acquired a conscience which is based both on a faith in the ultimate utilitarian value of any basic research and a feeling of responsibility to the country and to mankind.

In 1938 only 66 of the members and fellows of the Mineralogical Society worked in industry. At the present time this number has increased to 322. Only 10 years ago there were few if any x -ray diffraction units in petroleum geology research laboratories, yet today I do not think there are any in the United States without at least one and some have as many as a dozen.

The field of sedimentary geology was stimulated and enjoyed its greatest period of growth after it was accepted by the petroleum industry. Mineralogy and geochemistry, while accepted in principle, have yet to prove they can supply answers for practical application. The petroleum industry is eager and anxious to see this proven. In the 1957 AAPG Research Committee's article entitled "Research Needs in Petroleum Geology," of the 79 research problems listed at least 61 required the study of either or both mineralogy and geochemistry.

However, if we are going to take advantage of this opportunity it is essential that we become geologists, for the more geology we know the more mineralogy and geochemistry we will understand. Unfortunately, particularly in the field of sedimentary geology, the mineralogist-geochemist is often considered to work in a field remote from geology and to be incapable of determining the geologic significance of his own data. This is primarily the fault of the mineralogist-geochemist; too frequently little effort is made to understand the geologic history of the material being studied and only a limited effort is made to develop the geological significance of the data.

We are losing an immense amount of information and many, many badly needed ideas because we do not make enough effort to thoroughly understand the determining background of our data. In many cases the geologists have learned how to use mineralogy and geochemistry more effectively than the mineralogist and geochemist have learned how to

use geology. Perhaps we should call ourselves geomineralogists and thus make a greater effort to put some geology into mineralogy.

My personal feelings are that eventually we will be able to demonstrate that clay minerals and trace elements are more subtle environmental indicators than are fossils; such problems as diagenesis and the origin and migration of oil will be solved by combined clay mineral and geochemical studies; and that clay minerals and trace elements will develop into a major tool for finding stratigraphic traps. I believe we will see the time when most of the oil being found will be found by using mineralogical and geochemical data.

In conclusion, I wish to acknowledge the stimulating training I received at Pennsylvania State University from Drs. T. F. Bates, E. F. Osborn, and J. C. Griffiths, and in particular the inspirational teaching of Dr. P. D. Krynine, who first aroused my interest in the earth sciences. I owe a considerable debt to Dr. W. F. Bradley for the interest and advice he has willingly given me for the past ten years. Shell Oil Company and Shell Development Company have unhesitatingly provided me with a unique opportunity to pursue my field of study. A great many from Shell, but particularly Dr. R. A. Rowland, have provided the impetus and stimulation which have helped make my research such a consuming challenge

PROCEEDINGS OF THE THIRTY-NINTH ANNUAL MEETING
OF THE MINERALOGICAL SOCIETY OF AMERICA
AT ST. LOUIS, MISSOURI

C. S. HURLBUT, JR., *Secretary*.

The thirty-ninth meeting of the Society was held on November 6-8, 1958 at the Sheraton Jefferson Hotel, St. Louis, Missouri. A total of twenty-nine scientific sessions were held by the Geological Society of America and the affiliated Societies. Of these four were devoted to geochemistry, two to mineralogy and two to petrology, at which ninety-eight papers were presented. These represented over one-third of the total number of papers presented at the combined meetings.

The annual luncheon of the Society on November 7 was attended by 212 fellows, members and guests. Following the luncheon the sixteenth presentation of the Roebling Medal was made to Martin J. Buerger and the seventh presentation of the Mineralogical Society of America Award was made to Charles E. Weaver.

The Retiring President, George E. Goodspeed, addressed the Society on the afternoon of November 7. His subject was *Some Textural Features of Magmatic and Metasomatic Rocks*. The annual business meeting was held at 4:45 on the afternoon of November 7, at which brief reports were given by the Secretary, Treasurer and Editor. Professor M. J. Buerger gave an account of the organizational meeting of the International Mineralogical Association held in Madrid in April 1958.

COUNCIL ACTION

The 1958 Council of the Society met on the afternoon and evening of November 5 and on the afternoon of November 6 for a total of eleven hours. Twenty-six items of business were discussed.

Joint meeting with G.S.A. Council. For one hour on November 5 the Council met with the Council of the Geological Society of America to discuss problems of mutual interest. G.S.A. President Moore pointed out that their Council had called for a reduction in grants to associated Societies. A 10 per cent reduction had been suggested but no reduction would be made during the current year. However, M.S.A. may well look forward to diminishing grants. For the past few years G.S.A. has given M.S.A. \$8,000 a year to help pay the cost of the *American Mineralogist*.

Careers in Mineralogy. At the request of the 1957 Council, Dr. Charles Milton prepared a pamphlet on Careers in Mineralogy which was accepted by the Council. This pamphlet will be distributed by the American Geological Institute.

Articles for the Amateur. Since a considerable proportion of the membership of the Society is made up of nonprofessional mineralogists, it was proposed that occasional papers of popular appeal be printed in the *American Mineralogist*. The Council appointed Professor Richard H. Jahns to arrange for the preparation of such articles.

The 1959 Council met for three hours on November 8 and discussed the following items of business:

Treasurer. Earl Ingerson tendered his resignation as Treasurer of the Society effective November 15, 1958. This action resulted from his recent move from the U. S. Geological Survey in Washington to a post at the University of Texas. Dr. Ingerson has for the past eighteen years served the Society faithfully and well as its Treasurer. We all owe him a great debt of gratitude for his loyal and devoted service.

The Council appointed Miss Marjorie Hooker to fill the office of Treasurer until the next annual meeting in November 1959.

International Mineralogical Association. The Council voted that \$120.00 be paid during 1959 as the Society's dues to the International Mineralogical Association.

Honorary Fellows. Because the recent revision of the Constitution eliminated the category of Correspondents, the Council voted that the three remaining Correspondents be made Honorary Fellows. These are:

Pentti Eskola

L. J. Spencer

C. E. Tilley

The Council also voted to make H. J. Buttgenbach of Belgium a Honorary Fellow.

Committee Appointments. President Grim made various committee appointments which were approved by the Council and are printed below.

1959 COMMITTEES OF THE MINERALOGICAL SOCIETY OF AMERICA

Nominating Committee for Fellows

W. R. Foster, <i>Chairman</i>	1957-59
S. E. Clabaugh	1957-59
C. O. Hutton	1958-60
A. L. Anderson	1958-60
J. L. Kulp	1959-61
E. N. Cameron	1959-61

R. R. Franco
W. K. Gummer
R. A. Hoppin
H. S. Hill
W. D. Johnston, Jr.
A. A. Levinson
R. Rowland

Roebling Medal Committee

B. Mason, <i>Chairman</i>
M. J. Buerger
J. R. Goldsmith
H. F. McMurdie
A. Pabst
F. E. Wickman

Nominating Committee for Officers

W. F. Bradley, <i>Chairman</i>
G. C. Kennedy
E. W. Roedder
C. E. Weaver
H. Winchell

Program Committee

H. W. Fairbairn	1957-59
H. T. Evans	1957-59
O. R. Grawe	1958-60

MSA Award Committee

E. W. Nuffield, <i>Chairman</i>
J. W. Earley
A. F. Hagner
E. W. Heinrich
F. H. Stewart
O. F. Tuttle

Nomenclature Committee

C. Milton, <i>Chairman</i>	1957-59
H. Winchell	1957-59
C. S. Hurlbut	1958-60
M. Fleischer	1958-60
J. V. Smith	1959-61
C. Tunell	1959-61

Financial Advisory Committee

E. P. Henderson, <i>Chairman</i>	1959-61
S. S. Goldich	1959-60
A. Montgomery	1957-59

Membership Committee

D. J. Fisher, <i>Chairman</i>
P. B. Barton
H. Bannerman
C. W. Chesterman
G. E. Ericksen

Board of Associated Editors

J. Murdoch	1957-59
R. M. Garrels	1957-59
D. J. Fisher	1958-60
G. Faust	1958-60
G. W. Brindley	1959-61
A. Pabst	1959-61

Auditing Committee

M. E. Mrose, *Chairman*
G. Kullerud
E. W. Roedder

Mineralogical Abstracts

L. G. Berry, *Chairman*
Other members to be added by chairman.

Policy Committee

J. Murdoch, *Chairman*
S. S. Goldich
E. W. Nuffield
R. H. Jahns
C. Milton
M. Hooker, *ex officio*

Representatives

A.A.A.S.	C. Tolman	1957-1959
American Geological Institute		
Board of Directors	C. Frondel	1958-1959
	J. D. Barksdale	1959-1960
	R. M. Grogan	1960-1961
Government Relations Committee	P. F. Kerr	1957-1959
National Research Council	W. F. Bradley	1959-1961

International Mineralogical Association

Delegates

C. Frondel (voting delegate)
M. J. Buerger
E. Ingerson } alternates
T. S. Lovering }

Members of Commissions

Abstracts	H. Winchell
Data	G. T. Faust
New Minerals	D. McConnell
Museums	G. Switzer

REPORT OF THE SECRETARY

To the Council of the Mineralogical Society of America.

ELECTION OF OFFICERS AND FELLOWS

Six hundred and seventy-three ballots were cast in the election of officers, 184 by fellows and 503 by members. In addition 26 were irregular and are not included. The officers elected to serve in 1959 are:

President: Ralph E. Grim, University of Illinois, Urbana, Illinois.

Vice-President: Joseph Murdoch, University of California at Los Angeles, Los Angeles, California

Secretary: C. S. Hurlbut, Jr., Harvard University, Cambridge, Massachusetts.

Treasurer: Earl Ingerson, U. S. Geological Survey, Washington, D. C.

Editor: Lewis S. Ramsdell, University of Michigan, Ann Arbor, Michigan.

Councilors: (1959-1961) Edward W. Nuffield, University of Toronto, Toronto, Canada.

Wilfrid R. Foster, Ohio State University, Columbus, Ohio.

According to the provisions of the Constitution, the following have been elected to fellowship:

Philip Hauge Abelson, Geophysical Laboratory, Washington, D. C.

Henry N. Baumann, Jr., 1275 Wyoming Avenue, Niagara Falls, New York.

Margaret Fuller Boos, 2036 South Columbine, Denver, Colorado.

Joan Robinson Clark, 3 Pook's Hill Road, Bethesda, Maryland.

Theodor Ernst, University of Erlangen, Schlossgarten, Germany.

Sydney A. Forman, Chemical Division, Dept. of Agriculture, Ottawa, Ontario.

Albert Ronald Graham, 73 Greenland Road, Don Mills P.O., Ontario, Canada.

Claude Jean Guy Guillemin, 6 rue Pierre Bourdan, Paris 12e, France.

Djalma Guimaraes, Rua Gonçalves Dias, 3182, Belo Horizonte, Minas Gerais, Brazil.

Gunnar Kullerud, Geophysical Laboratory, Washington, D. C.

Jan Kutina, Dept. of Geochemistry, Charles University, Albertov 6, Prague II, Czechoslovakia.

Alfred A. Levinson, Central Laboratory, The Dow Chemical Co., Freeport, Texas.

Ivan Herbert Milne, 832 Twelfth St., Oakmont, Pennsylvania.

Herbert O'Daniel, Mineralogisches Institut der Universität, Senckenberg Anlage 30, Frankfurt, Germany.

Robert George Schmidt, U. S. Geological Survey, Washington, D. C.

Ming Shan Sun, Box 26, Campus Station, Socorro, New Mexico.

Karl Hans Wedepohl, Mineralogisch Institut der Universität, Göttingen, Germany.

Kenzo Yagi, Institute of Earth Sciences, Tohoku University, Tomizawa, Sendai, Japan.

Revision of the Constitution

The following question appeared on the ballot sent to members and fellows in September 1958, "Shall the changes in the Constitution and By-Laws as printed in the *American Mineralogist* 43: pp. 611-614, 1958 be accepted?" Of the 617 fellows and members voting on this question, there were 585 affirmative votes and 32 negative votes.

MEMBERSHIP STATISTICS

November 1, 1958

	1957	1958	Gain	Loss
Correspondents.....	3	3	0	0
Fellows.....	356	360	18	14
Members.....	1,075	1,185	193	83
Subscribers.....	1,041	1,195	236	82

The above figures show a net gain of 4 fellows, 110 members and 154 subscribers. Considering the four groups together, there is a gain of 268 giving a total of 2743.

During 1958, the Society lost through death seven fellows: G. H. Anderson, New York, N. Y.; Oliver Bowles, Washington, D. C.; Carl Faessler, Quebec, Canada; F. F. Grout, Minneapolis, Minnesota; E. C. Jacobs, Burlington, Vermont; M. H. Stow, Lexington, Virginia; A. W. Winchell, Hamden, Connecticut.

Respectfully submitted,

C. S. HURLBUT, JR., *Secretary*

REPORT OF THE TREASURER FOR 1958

To the Council of the Mineralogical Society of America:

Your Treasurer submits herewith his report for the fiscal year beginning August 1, 1957, and ending July 31, 1958.

RECEIPTS

Dues and subscriptions.....	\$13,756.85
Sale of back numbers.....	2,773.77
Authors' charges on reprints.....	3,759.06
Income from endowment.....	6,593.32
Geological Society of America aid for printing the Journal.....	8,000.00
Advertising.....	864.76
Sale of Index, vols. 31-40.....	409.40
Sale of Index, vols. 21-30.....	73.15
Sale of Index, vols. 1-20.....	38.70
N.S.F. grant for printing the Index to vols. 31-40.....	4,500.00
Harvard University grant toward printing Harvard issue.....	1,539.00
Sale of securities.....	1,312.50
Credit held in overpayment of account.....	7.00
	<hr/>
	\$43,627.51
Cash on hand, August 1, 1957.....	1,531.53
	<hr/>
	\$45,159.04

DISBURSEMENTS

Printing and distribution of the Journal (6 issues).....	\$19,801.06
Printing and distribution of reprints.....	3,213.07
Honoraria to Editor, Secretary and Treasurer.....	1,250.00
Clerical assistance.....	1,266.65
Printing and stationery.....	330.29
Program and abstracts (1957).....	286.75
Postage and express.....	550.32
Expenses of officers to 1957 meeting.....	25.00
New securities purchased.....	10,157.57
Commission on new securities.....	110.26
Tax on new securities.....	4.67
Office supplies.....	63.45
Safety deposit box.....	7.15
Refunds on subscriptions.....	25.55
Telephone and telegrams.....	4.10
Committee expenses.....	4.16
Roebing Medal (1957).....	148.23
Mineralogical Society of America Award.....	7.00
Contribution to American Geological Institute.....	600.00
Contribution to Mineralogical Abstracts.....	350.00
Contribution to International Mineralogical Assoc.....	120.00
Printing Index, vols. 31-40.....	4,679.49
Refund to N.S.F. on Index grant.....	323.40
Refund of dividend on stock sold.....	20.00
Printing micro text cards.....	198.73
Checks returned.....	13.00
	<hr/>
	\$43,559.90
Cash on hand, August 1, 1958.....	1,599.14
	<hr/>
	\$45,159.04

The endowment funds of the Society as of July 31, 1958, consist of the following securities:

BONDS

6M Atlantic Coast Line, 4½	\$ 5,257.50
5M New York Central, 5	4,300.00
5M Southern Railway, 5	5,743.75
	<hr/>
	\$ 15,301.25

PREFERRED STOCKS

500 Union Pacific, 4	\$ 4,570.25
200 Southern California Edison, 4.88	5,250.00
150 Penn-Texas, 1.60	4,487.50
60 Jones and Laughlin, A, 5	4,987.50
55 United States Steel, 7	6,946.20
50 Virginia Electric and Power, 5	5,942.50
40 Potomac Electric Power, 2.44	2,000.00
10 Consolidated Edison, 5	1,066.64
	<hr/>
	\$ 35,250.59

COMMON STOCKS

346 Potomac Electric Power	\$ 4,966.73
215 Columbia Gas	3,436.25
200 Greyhound Corporation	2,300.00
175 Standard Oil of New Jersey	1,761.65
150 Consolidated Denison	3,096.00
100 Buckeye Steel	3,800.00
100 Columbus and Southern Ohio Electric	2,087.50
100 Diana Stores	1,250.00
100 Pittsburgh and West Virginia	2,787.50
100 Plymouth Cordage	4,750.00
100 Southern Natural Gas	3,600.00
87 Kroger Company	2,034.03
83 American Telephone and Telegraph	11,255.98
60 United Fruit	3,067.50
50 Boston Insurance	1,500.00
50 Chesapeake and Ohio	2,368.75
50 Phelps Dodge	1,975.00
50 Public Service Electric and Gas	1,586.40
50 Sinclair Oil	2,968.75
50 Spencer Kellogg	1,258.33
30 U. S. Playing Card	2,411.25
	<hr/>
	\$ 64,261.62
	<hr/>
	\$114,813.46

Respectfully submitted,
EARL INGERSON, *Treasurer*

REPORT OF THE AUDITING COMMITTEE

To the President of the Mineralogical Society of America:

The Auditing Committee has examined and verified the accounts of the Treasurer of the Mineralogical Society of America for the fiscal year beginning August 1, 1957, and ending July 31, 1958. The securities listed in the Treasurer's report, with all future coupons on the coupon bonds attached, are in the safety deposit box at the 17th and G Street Branch of the Riggs National Bank in Washington, D. C.

Respectfully submitted,
FRANCIS R. BOYD, JR.
MARY E. MROSE
WILLIAM T. PECORA, *Chairman*

REPORT OF THE EDITOR FOR 1958

The 1958 volume of *The American Mineralogist* contained 1247 pages, as compared with 940 and 974 for 1957 and 1956 respectively. This is an increase of about $30 \pm$. The 111 articles printed were received from 192 authors. A wide geographic distribution is shown by the fact that these authors represented 20 universities, 11 governmental agencies and 4 industrial laboratories in the United States, and 16 universities and 3 government agencies from outside the United States. This gives a total of 54 institutions, which is an increase of nearly 50% over 1957.

In spite of the increase in size, the backlog of manuscripts on hand is larger than that of a year ago, so continued larger issues will be necessary in 1959.

Five new minerals were described in detail for the first time in Volume 43. There were 25 book reviews. The section on New Mineral Names, New Data and Discredited Minerals, so ably conducted by Dr. M. Fleischer, contains 79 entries. This is an essential and valuable section, and we owe a great deal to Dr. Fleischer for this service.

As in former years, the Geological Society of America has continued its liberal financial support toward the cost of printing the Journal, and we are grateful for this assistance.

The quality of the Journal depends in large part upon the advice of the associate editors and others who read the manuscripts submitted for publication. The fields covered by our Journal are so varied that it is important that we use a wide range of talent from our memberships. The editor wishes to thank the Associate Editors, Drs. C. L. Christ, E. Wm. Heinrich, Robert Garrels, Joseph Murdoch, George T. Faust and D. Jerome Fisher, and likewise the following who have aided in reading manuscripts: Drs. W. F. Bradley, G. W. Brindley, R. M. Denning, H. T. Evans, Jr., A. Pabst, J. P. Smith and H. Winchell. The editor also wishes to express his appreciation to the George Banta Company Inc., publisher of our Journal. The excellent appearance of the Journal, and the fine cooperation received are deeply appreciated.

A summary of the contents of Vol. 43 is given in the following table.

DISTRIBUTION OF SUBJECT MATTER IN VOLUME 43

Subjects	Number	Pages	% of Total
Leading articles.....	70	940	79.7
Short articles.....	41	106	9.0
		1,046	88.7
Memorials.....	7	37	
Awards.....	2	15	
Proceedings of the Society.....	1	18	
		70	6.0
Book Reviews.....	25	20	
New Mineral Names.....	79	34	
Notes.....	19	10	
		64	5.3
Total.....		1,180	100.0
Membership list.....		45	
Index, Title Page, Table of Contents.....		22	
		67	
Grand Total.....		1,247	

Respectfully submitted,
LEWIS S. RAMSDELL, *Editor*

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Annual Meeting Places

- | | |
|-------------------------------|--------------------------------|
| 1920 Chicago, Illinois | 1940 Austin, Texas |
| 1921 Amherst, Massachusetts | 1941 Boston, Massachusetts |
| 1922 Ann Arbor, Michigan | 1942 No meeting held |
| 1923 Washington, D. C. | 1943 No meeting held |
| 1924 Ithaca, New York | 1944 No meeting held |
| 1925 New Haven, Connecticut | 1945 Pittsburgh, Pennsylvania |
| 1926 Madison, Wisconsin | 1946 Chicago, Illinois |
| 1927 Cleveland, Ohio | 1947 Ottawa, Canada |
| 1928 New York, N. Y. | 1948 New York, N. Y. |
| 1929 Washington, D. C. | 1949 El Paso, Texas |
| 1930 Toronto, Canada | 1950 Washington, D. C. |
| 1931 Tulsa, Oklahoma | 1951 Detroit, Michigan |
| 1932 Cambridge, Massachusetts | 1952 Boston, Massachusetts |
| 1933 Chicago, Illinois | 1953 Toronto, Canada |
| 1934 Rochester, New York | 1954 Los Angeles, California |
| 1935 New York, N. Y. | 1955 New Orleans, Louisiana |
| 1936 Cincinnati, Ohio | 1956 Minneapolis, Minnesota |
| 1937 Washington, D. C. | 1957 Atlantic City, New Jersey |
| 1938 New York, N. Y. | 1958 St. Louis, Missouri |
| 1939 Minneapolis, Minnesota | |

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1958 ANNUAL MEETINGS AT ST. LOUIS, MISSOURI

The meetings of the following Societies were held on November 6-8, 1958.

The Geological Society of America—71st
 The Paleontological Society—50th
 The Mineralogical Society of America—39th
 Society of Economic Geologists—39th
 Geochemical Society—3rd
 Association of Geology Teachers

The complete program listed titles and abstracts of 323 papers. Of these 161 seemed to be of special interest to members of The Mineralogical Society of America. Authors and titles of these papers are listed below. Since by Council action the abstracts will not be reprinted in *The American Mineralogist*, references are given to the *Bulletin* of the Geological Society of America, Vol. 69, Number 12, Part 2, December 1958, pp. 1527-1770, where the abstracts of all the papers of the various societies are reprinted, arranged alphabetically by authors.

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NOTES AND NEWS

MICROFRAMEWORKS*

ELIZABETH A. WOOD, *Bell Telephone Laboratories, Murray Hill, N. J.*

The purpose of this paper is to call attention to those regular, repetitive features of crystalline bodies whose repeat distances are in general larger than the interatomic distances of the crystal structure, but not visible to the unaided eye. These I shall call *microframeworks*.

One of the reasons for interest in them is that they may result in the development of a new area of investigation in physics and chemistry since the properties of substances under the conditions that exist in some microframeworks will be different from their bulk properties. Secondly, as a result of this difference in properties and also because of their very small size and precise dimensions, microframeworks may be of technological interest.

These two points will be illustrated under the headings of several types of microframeworks: *micromolds*, *microsandwiches*, *microgrids*, *microcages*, *microsieves*, and *microjigs*.

1. *Micromolds*

Chrysotile asbestos has been described (Bates, Sand, and Mink, 1950) as composed of closely packed tubes. Noll and Kircher (1950) show electron micrographs of cross sections of these tubes and state that their outside diameter is 150–300 Å. Such a structure could serve as a precisely dimensioned tubular micromold whose tubes could be filled with the substance whose properties were to be studied.

Consider first magnetic properties. It is well known that the domain-wall energy of a ferromagnetic material is such that a wall cannot exist in a particle of about 200 Å diameter (Kittel, 1946). Such a particle, consisting of a single domain, has a high coercive field. A body made up of such particles, packed in a matrix would therefore be an excellent permanent magnet. If the particles consisted of parallel rods, the effect of shape anisotropy would be added and the body would be a better permanent magnet than any now known.

Williams and Sherwood (1958) attempted to fill chrysotile tubes with various magnetic materials and achieved a 600-fold increase of the coercive force in cobalt compared to that of the bulk material, but this is far short of what was expected from theory. This unsatisfactory result may support the conclusion of F. L. Pundsack (1956) whose density measure-

* The substance of this paper was the subject of the retiring president's talk at the banquet of the American Crystallographic Association meeting in Milwaukee, June 26, 1958.

ments indicate that the supposed tubes of chrysotile asbestos are in fact filled with solid material. If this was true of the specimen which we tried to fill with the magnetic material, the volume filled could only have been the interstices between the rods, giving an inadequate amount of magnetic material present in the composite body.

If a tubular micromold was filled with a metal, the resistance of the microwires would be high because of their small diameter and a new type of resistor would result (Doucette, 1958). Since the "wires" would maintain their shape and size in spite of phase transformations, including melting, the changes of conductivity with changes of phase could be studied.

It is well known that metals in thin films do not have the same density, optical constants, and electric constants as bulk metals (Lewis, 1950) and that the epitaxial influence of the substrate may even result in a change of crystal structure of the deposited substance (Pashley, 1956). The study of properties of materials in micromolds is a new field of investigation where one may expect the discovery of new properties of materials. Since the material in a micromold is not in contact with air except for a negligibly small area, reactive materials can be investigated without the necessity of a controlled atmosphere. Although the tubular micromolds so far described have been crystals with spaces available for filling, microframeworks may also exist in which rod-like inclusions of a material different from the matrix may be selectively removed or replaced with the same end effect as that of a tubular micromold.

In addition to tubular or "one-dimensional" micromolds, lamellar or "two-dimensional" micromolds may be used. A filled lamellar micromold has the properties of a *microsandwich* which will be discussed below. A three-dimensional micromold would probably have to be filled as the crystal grew. The *microcage* discussed below is related to this.

2. *Microsandwiches*

Microsandwiches are layered structures on a very small scale whose adjacent layers differ from each other in composition or structure. Such microframeworks are called polycrystals (Donnay and Donnay, 1953) and may be composed of such different substances as pyrite and uraninite (King, 1957). In current microwave technology, components with closely-spaced thin metallic filaments have been needed (Reed, 1955). Microsandwiches with metallic layers, either originally present or introduced by selective replacement might find application in this field.

3. *Microgrids*

The construction of ruled gratings for optical diffraction is an exacting mechanical task. Microridges on crystal surfaces might, in favorable

cases, serve as an optical grating or as a mold for a replica that could be used for an optical grating as suggested by Farney (1958). Slip planes, as in twinned calcite, produce such a ridged surface. Alternating development of faces, as on the cube face of pyrite and the prism face of quartz, if sufficiently regular, might provide a suitable microgrid. In some crystals, e.g., albite, polysynthetic twinning results in twin lamellae of uniform width. Selective etching of these would produce a microgrid.

The remaining three categories comprise small scale microframeworks, examples of which are already in use.

4. *Microcages*

A microcage is a crystal microframework that is used to trap something that would otherwise escape. The clathrate compounds are an example. In the quinol-rare gas clathrates, the gas atoms are mechanically trapped in quinol cages as close together as they would be if the gas were under a pressure of 91 atmospheres. The oil industry uses clathrate compounds to trap selectively petroleum derivatives whose boiling points are not sufficiently different from each other to allow separation by fractional distillation (Banerjee, 1958).

5. *Microsieves*

A microsieve is a crystal microframework which, because of the shape or size of openings in it, is useful for separating one substance from another. An example is the zeolites which have large (ca. 10 Å) cavities interconnected by smaller passages (ca. 4 Å in diameter). These have recently been used to separate straight chain hydrocarbons from branched chain hydrocarbons (Breck, 1958). They have also been called "molecular sieves."

6. *Microjigs*

A microjig is a crystal microframework in which something is held for processing in some way. Examples are the urea and thiourea crystals used by Brown and White (1958) to hold monomers end to end during electron-irradiation which results in their linkage into straight chain polymers.

The list of microframeworks given in this paper is not exhaustive. Others will be discovered or produced synthetically. It is hoped that mineralogists and crystallographers will pay special attention to potential microframeworks in the crystals they find or grow. These, in the hands of engineers, may make possible the construction of useful devices on a scale that would be prohibitively difficult for manufacturing processes. In the hands of crystallographers and physicists, microframeworks may be a new tool for the study of solids.

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SYNTHESIS OF SABUGALITE¹

GEORGE B. MAGIN, JR., GEORGE J. JANSEN,² AND BETSY LEVIN,
U. S. Geological Survey, Washington 25, D. C.

In his paper on sabugalite, $\text{HAl}(\text{UO}_2)_4(\text{PO}_4)_4 \cdot 16\text{H}_2\text{O}$, Frondel (1951) stated that his efforts to synthesize this mineral had been unsuccessful.

¹ Publication authorized by the Director, U. S. Geological Survey.

² Present address, Republic Steel Corporation, Cleveland, Ohio.

Two of the authors (Magin and Jansen) were at that time working on the synthesis of uranyl phosphates and successfully synthesized this mineral. The method used was similar to that used by Beintema (1938) in his synthesis of uranocircite. The stoichiometric amount of uranyl acetate dihydrate (200.0 mg) was added to 27.5 mg. of aluminum chloride hexahydrate in a large volume of water (600 ml.). A threefold excess of phosphoric acid (1.4 ml. of a 100 g/liter solution of H_3PO_4) was added. A yellow fine-grained precipitate formed slowly and was digested in a covered beaker on the steam bath for 2 days. The chemical analysis of the resulting product is given in Table 1.

Because the synthetic crystals were exceedingly fine grained, they did not yield satisfactory optical data. However, the average index of refraction of the synthetic product indicated a value of about 1.57, which is within the range of indices given by Frondel for natural sabugalite with water content approximately the same as that of the synthetic product.

Both the natural and the synthetic sabugalite were indexed on the basis of the data given by Frondel (1951)—tetragonal ($I4/mmm$), $a_0=6.96 \text{ \AA}$ and $c_0=19.22 \text{ \AA}$. The results obtained with the use of these data are not satisfactory, however, as not all the reflections could be indexed. Good single crystal patterns could not be obtained due to the very thin and warped nature of the crystals. Until better material is discovered, assignment of the space group to which this mineral belongs must be only tentative.

In Table 2 Frondel's powder data of sabugalite are given for comparison with our data. The line at $d=4.86 \text{ \AA}$ has been resolved on our patterns of both the synthetic and the naturally occurring material, to $d=4.93 \text{ \AA}$ and 4.80 \AA , which then can be indexed as (110) and (004) respectively.

The writers gratefully acknowledge the help of Mrs. Daphne R. Ross

TABLE 1. CHEMICAL ANALYSIS (IN PER CENT) OF SABUGALITE

	Synthetic (G. B. Magin, Jr., Analyst)	Natural (Frondel, 1951)	Theoretical
UO_3	64.43	65.22	64.41
Al_2O_3	2.90	2.65	2.87
P_2O_5	15.75	16.08	15.99
H_2O	16.67	15.93	16.73
Total	99.75	99.88	100.00

TABLE 2. X-RAY POWDER DATA FOR NATURAL AND SYNTHETIC
SABUGALITE: $\text{HAl}(\text{UO}_2)_4(\text{PO}_4)_4 \cdot 16\text{H}_2\text{O}$

Indexed on tetragonal unit cell: space group $I4/mmm$; $a_0 = 6.96 \text{ \AA}$, $c_0 = 19.22 \text{ \AA}$; $\text{CuK}\alpha$ radiation ($\lambda = 1.5418 \text{ \AA}$); nickel filter; camera diameter = 114.59 mm.; cut-off at 11.0 \AA

hkl	$d_{\text{calc.}}$	Synthetic sabugalite (Film No. 9899)		Natural sabugalite (Film No. 8991) Locality: Mina da Quarta Seira, Sabu- gal, Portugal		Natural sabugalite (Fron del, 1951) Locality: Mina da Quarta Seira, Sa- bugal, Portugal	
		$d_{\text{meas.}}$	I	$d_{\text{meas.}}$	I	$d_{\text{meas.}}$	I
002	9.605	9.61	100	9.61	100	9.69	10
		9.03*	9	8.93*	9		
		8.43	1	8.43†	9		
		6.71	3	6.76	3		
101	6.544	6.37	3	6.37	4	6.56	1
				5.44†	3	5.59	1
		5.31	4				
110	4.922	4.93	18	4.93	35		
004	4.802	4.80	18	4.80	35	4.86	9
103	4.713						
112	4.380	4.55	25	4.58	13		
		4.40	9	4.40	9	4.39	4
		4.17(<i>vb</i>)	2	4.19(<i>b</i>)	2		
		3.97	2	3.63†	4		
200	3.480	3.48	35	3.48†	50	3.47	8
114	3.437						
105	3.364	3.36	9	3.36	13	3.36	1
006	3.202	3.22(<i>b</i>)	9	3.22(<i>b</i>)	9	3.22	$\frac{1}{2}$
211	3.073	3.05(<i>b</i>)	2	3.08	2	3.06	$\frac{1}{2}$
		2.94	3	2.93	6		
		2.88	3				
204	2.818					2.818	1
213	2.800						
116	2.684						
		2.63	6	2.62	4		
107	2.553						
220	2.461	2.46(<i>b</i>)	4	2.47(<i>b</i>)	4	2.452	2
214	2.418						
008	2.401	2.40	6	2.40	9	2.389	2

b = broad. *vb* = very broad.

* Possibly a lower hydration state of sabugalite.

† Meta-autunite.

TABLE 2 (Continued)

<i>hkl</i>	<i>d</i> _{calc.}	Synthetic sabugalite (Film No. 9899)		Natural sabugalite (Film No. 8991) Locality: Mina da Quarta Seira, Sabu- gal, Portugal		Natural sabugalite (Fron del, 1951) Locality: Mina da Quarta Seira, Sa- bugal, Portugal	
		<i>d</i> _{meas.}	I	<i>d</i> _{meas.}	I	<i>d</i> _{meas.}	I
222	2.384						
206	2.356						
301	2.303	2.29	2	2.30	2		
		2.24	2	2.25	1		
310	2.201						
224	2.190	2.19(<i>b</i>)	9	2.20(<i>b</i>)	13	2.188	6
303	2.182						
118	2.165						
312	2.146						
217	2.058	2.08(<i>vb</i>)	3	2.10(<i>vb</i>)	4		
109	2.040						
		1.989	2	1.998	1		
		1.918	4	1.926	6		
		1.859	3	1.863	3		
		1.791(<i>b</i>)	3	1.794(<i>vb</i>)	4	1.792	1
		1.740	4	1.743	9	1.726	1
		1.670	2	1.684	1	1.641	$\frac{1}{2}$
		1.603	1	1.605	2		
		1.578	2	1.583	3		
		1.551	2	1.556	2	1.552	1
		1.523	2	1.528	3		
		1.421(<i>b</i>)	1	1.423(<i>b</i>)	2		
		1.366(<i>b</i>)	3	1.370(<i>vb</i>)	4	1.364	1

of the U. S. Geological Survey in the *x*-ray studies. This work is part of a program being conducted by the U. S. Geological Survey on behalf of the Division of Research of the U. S. Atomic Energy Commission.

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DIRECTIONAL GRINDING HARDNESS OF QUARTZ
BY PERIPHERAL GRINDING*R. M. DENNING AND M. A. CONRAD, *University of Michigan,
Ann Arbor, Michigan.*

Recently Giardini (1) and Giardini and Conrad (2) have successfully studied the directional grinding hardness of strontium titanate and silicon single crystals by cutting oriented disks and grinding the periphery of the disks in a lathe. An outstanding feature of the method is that information on the symmetry of hardness vectors is quickly obtained from the measurements. It occurred to the writers that the peripheral grinding method might be used to verify experimentally the hypothesis that grinding hardness is a non-centrosymmetric property of crystals. Quartz was chosen for the study as being most suitable from the standpoint of availability and mechanical properties, among non-centrosymmetric crystals. At the outset, quartz is not ideal for the purpose because of the near isotropy of its other physical properties. Quartz possesses a poor to distinct prismatic cleavage $\{10\cdot0\}$ and a similar pair of rhombohedral cleavages $\{10\cdot1\}$ and $\{\bar{1}0\cdot1\}$. The cleavage of quartz may be readily observed on a rough ground single crystal sphere. Reflection maxima are easily detected so that such a sphere can be accurately oriented in reflected light; the only ambiguity is that introduced by the inability of the observer to distinguish between $\{10\cdot1\}$ and $\{\bar{1}0\cdot1\}$. The presence of cleavage suggests a likelihood of measurable hardness variation, although the large number of cleavage planes (nine) indicates that the variations may not be very large.

The reader is referred to the paper by Giardini and Conrad (2) for a description of the method of peripheral grinding. External morphology and etch figures were used as an aid in selecting single crystal oriented disks in this study. The hand of the crystals was determined by light figures from etched surfaces and the rotation of the plane of polarization of light traveling parallel to the optic axis as described by Parish and Gordon (3).

The disks were mounted on a spindle and centered as well as possible in the lathe chuck. The chuck was divided into five degree increments which could be read at a fixed index. The centering and radial reductions were determined by means of a dial indicator. Thus the reduction of the

* Contribution No. 225 from the Department of Mineralogy, University of Michigan, Ann Arbor, Michigan. This investigation was supported financially by the Office of Naval Research.

radius of the disk due to grinding could be recorded for the desired angular position of the chuck (azimuth). The lathe cross feed served to locate the dial indicator and also to locate the abrasive tape against the disk for grinding.

The indicator was graduated to 0.001 inch. Readings to 0.0001 inch were estimated and are believed to be correct to 0.0003 inch. The reading of the azimuth angle is not particularly critical, but is correct to within one degree. The thickness and maximum and minimum disk diameters were measured with a micrometer graduated to 0.001 inch. Experimental conditions were essentially the same as those used in the silicon and strontium titanate studies previously mentioned.

Three disk orientations were chosen for the present study: a [00·1] disk and a disk of general orientation (zone axis; $\phi = 226^\circ$, $\rho = 37^\circ$) to investigate the non-centrosymmetric nature of the hardness and a [01·0] disk as an example of two-fold symmetry.

As a test of the validity of the method, an optically isotropic pyrex disk was subjected to a series of measurements. The initial eccentricity due to lack of perfect centering was reduced during the first grinding period. Observations after further grinding indicated that the equilibrium between eccentricity and hardness attained during the first grinding period was valid within experimental error. Thus it is reasonable to conclude that any initial eccentricity will be reduced during grinding and that variation in hardness will be the principal factor upon which differences in radial reduction will be dependent.

The reproducibility of the method was checked by making observations of the radial reductions produced by successive grinding increments. In the case of quartz, the observed relative grinding hardnesses produced by successive tests agree within 10 per cent.

Relative grinding hardness values were computed from the observed radial reductions using the following definition for relative grinding hardness (4): $RGH = K_s/K_n$ where K_s is the grinding constant for a direction of arbitrarily chosen unit hardness and K_n the grinding constant in the direction under consideration. The grinding constant $K = v/FT$, where v is the volume removed, F is the force normal to the grinding surface and T is the time of grinding.

Hardness values were computed for each ten degree increment using the largest radial reduction as the standard value for each test. Successive values for each orientation were then averaged and the standard deviation was computed.

Average relative hardness values for each disk were then determined with the above equations, using the volume removed during the entire grinding period to determine the grinding constant (K) for each disk.

provement in technique probably would permit a reproducible hardness curve for this orientation.

The two-fold symmetry of the hardness curve of the $[01\cdot0]$ disk is apparent. The curve for the disk of general orientation does not approach the two-fold symmetry so closely as the former. It is apparent

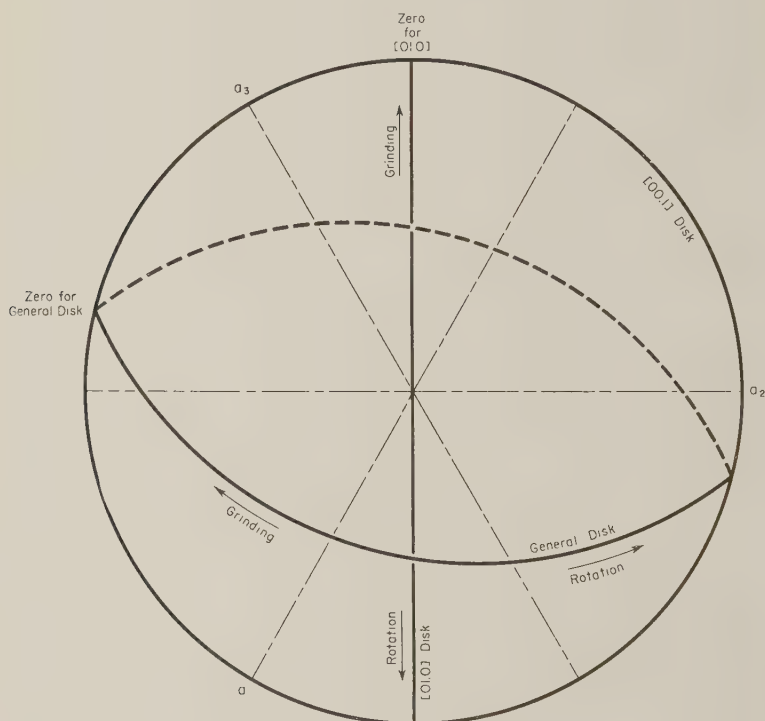


FIG. 2. Cyclographic projection of the quartz disks.

that the slight deviation from two-fold symmetry may be fortuitous and cannot be considered as significant.

It is not possible to combine the curves obtained from a series of disks to form a useful and meaningful solid of relative grinding hardness, since grinding hardness is dependent upon the orientation of the plane being ground and the direction within the plane.

It is well to consider the symmetry that the two-dimensional hardness curves would be expected to show for the possible symmetry elements of crystals. A disk whose axis coincides with an n -fold symmetry axis of rotation or rotary reflection must exhibit a hardness curve of n -fold or compatible symmetry. If the symmetry axis is an axis of rotary inversion, the relations (which can be verified easily with the aid of a stereo-

graphic projection) are as follows: if the disk axis be parallel to $\bar{2}$, $\bar{3}$, $\bar{4}$, or $\bar{6}$, the hardness variation shown by peripheral grinding of the disk must be compatible with 1, 6, 4, or 3-fold symmetry respectively.

If a disk is cut from a centrosymmetric crystal the least symmetry which can be manifest is twofold. Incidentally an important point, about which there has been some confusion, is the equivalence of hardness vectors required by a center of symmetry on sawed (or cleaved) crystal surfaces. Figure 3 illustrates a crystal which has been sawed on the stippled plane. A hardness vector H on the sawed surface of Part A is repeated by a center of symmetry as H' on Part B.

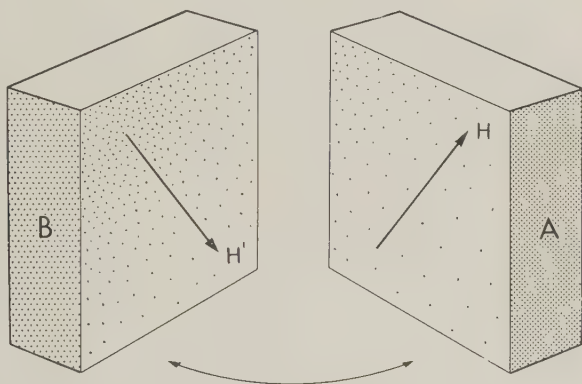


FIG. 3. Repetition of a hardness vector in a cleaved centrosymmetric crystal.

A symmetry plane parallel to the axis of the disk will reveal itself only if curves are obtained for opposite directions of disk rotation. Under such conditions the two curves will be related to each other by a line of reflection parallel to the trace of the symmetry plane.

Ideally (from symmetry considerations alone) if a sufficient number of properly chosen disks be ground, some in opposite directions of rotation, a unique crystal class determination should be possible from hardness measurements alone. Furthermore, it should be possible to distinguish between right and left enantiomorphs. Practically, this has not been accomplished. About the best that has been done is to show measurements consistent with the symmetry elements of the substance investigated.

Anisotropy (A) may be expressed by the following formula, in which H is relative grinding hardness and R is radial reduction:

$$A = \frac{2|H_1 - H_2|}{H_1 + H_2} \text{ or } \frac{2|R_1 - R_2|}{R_1 + R_2}$$

Using the maximum and minimum hardness values, quartz has an anisotropy of 0.25 for the $[01\cdot0]$ disk, 0.24 for the disk of general orientation, and 0.11 for the $[00\cdot1]$ disk. As a comparison, the anisotropy of relative grinding hardness of diamond (4) is two which is, of course, the theoretical maximum anisotropy defined by the above equation. The anisotropy of hardness of strontium titanate is 0.70 (1) and of silicon 0.26 (2). No single crystal as yet studied has displayed isotropic grinding hardness (anisotropy of zero).

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TEPHROITE FROM CLARK PENINSULA, WILKES LAND, ANTARCTICA

BRIAN MASON, *The American Museum of Natural History, New York.*

In January 1957 a United States base (Wilkes Station) was established on Clark Peninsula (about 66° S, 110° E), in Wilkes Land, in connection with the International Geophysical Year. Mr. Walter Sullivan accompanied the expedition as a reporter for the *New York Times*, and while ashore at Wilkes Station he noticed a vein of black rock with a metallic sheen in the light-colored gneisses forming the bedrock of this region. He observed that the vein outcropped at several points on a ridge near the site of Wilkes Station, and that where it was exposed it was less than a yard wide. He also noticed green stains of malachite along joints in the country rocks. Mr. Sullivan collected a number of specimens and on his return to New York he gave them to the museum. Laboratory examination showed that the black vein material consisted largely of the manganese silicate tephroite. Since tephroite is not a common mineral and has not been previously recorded from Antarctica the material has been investigated in some detail.

The specimens as collected are coated with a black iridescent film with a metallic luster, evidently a manganese oxide. On fresh fracture, however, the material has a dark ash-gray color typical of tephroite; small grains of yellow spessartite are scattered through the tephroite, and occasional small patches of white barite and pink rhodonite are present.

The tephroite is granular, the individual grains being up to 3 mm. in diameter.

In thin section the rock is seen to be a granoblastic aggregate of tephroite, spessartite, rhodonite, and barite, of which tephroite makes up more than 90%. A small amount of black and reddish-brown alteration products is present along cracks. The spessartite has a very faint yellowish tinge and is completely isotropic. The tephroite and the rhodonite have a somewhat grayish tint in comparison to the spessartite; they are very similar in appearance, but the more pronounced cleavage and the much lower birefringence of the rhodonite aid in distinguishing it. The tephroite and the rhodonite contain numerous very small opaque inclusions. A few grains of a brown anisotropic mineral with very high refractive index and straight extinction were seen in the thin section, generally in association with the spessartite; they are probably rutile.

The rhodonite has refractive indices $\alpha=1.721$, $\beta=1.725$, $\gamma=1.732$, very similar to those of a rhodonite from Harstigen, Sweden, described by Sundius (1931). These indices suggest a composition in which about 20% of the manganese is replaced by calcium, iron, and magnesium. The spessartite has refractive index $n=1.804$ and unit cell dimension $a=11.68 \text{ \AA}$. The diagrams of Sriramadas (1957) show that these properties correspond to a composition of 84% spessartite component, 9% andradite, and 7% grossularite.

In immersion liquids the grains of tephroite are generally colorless and transparent, but a few are stained yellow or brown by alteration products. Thick grains show weak pleochroism, X=pale yellow, Z=pale blue. The refractive indices are $\alpha=1.770$, $\beta=1.803$, $\gamma=1.819$. It is optically negative, $2V=71^\circ$ (determined on the universal stage). The density of the tephroite is 4.08, and its hardness is $6\frac{1}{2}$. The luster is vitreous, inclining to greasy, and the streak is pale gray-brown.

A sample of the purest material was carefully crushed and the tephroite selected by hand-picking. Rhodonite and barite were readily removed in this way, but it was not possible to prepare a sample completely free from spessartite. The analyzed material contains about 2% of spessartite. The chemical analysis was made by Dr. H. B. Wiik, and the results are as follows: SiO_2 29.31, TiO_2 0.05, Al_2O_3 0.48, Fe_2O_3 1.21, FeO 1.67, MnO 65.62, MgO 0.71, CaO 0.70, Na_2O 0.07, K_2O 0.00, P_2O_5 0.07, $\text{H}_2\text{O}+$ 0.19, $\text{H}_2\text{O}-$ 0.02, CO_2 0.00, total 100.10. The Al_2O_3 percentage corresponds to the amount of spessartite observed in the analyzed material. If allowance is made for this, the molecular proportions of $\text{MnO}+\text{MgO}+\text{FeO}+\text{CaO}:\text{SiO}_2$ is 0.9643:0.4720, or 2.00:0.98, in agreement with the formula Mn_2SiO_4 for tephroite. In this tephroite there is little replacement of manganese by other elements, it containing over 94% of the Mn_2SiO_4 component. In this respect it is comparable to

the tephroite from Franklin, New Jersey, described by Palache (1928).

The lack of information regarding the local geology at Clark Peninsula makes it difficult to discuss the paragenesis of these manganese silicate minerals. The country rock, judging from the specimens which Mr. Sullivan collected, is a coarse-grained quartz-andesine-biotite-almandine-cordierite-sillimanite gneiss. This association is typical of high-grade regional metamorphism. The assemblage tephroite-spessartite-rhodonite is compatible with crystallization at medium to high temperatures; this assemblage is stable at the liquidus in the system $\text{MnO-Al}_2\text{O}_3\text{-SiO}_2$ (Snow, 1943). Mr. Sullivan describes the occurrence of the manganese silicates as a vein; it is also conceivable that it may be a narrow band of originally manganese-rich material which has been subjected to regional metamorphism along with the country rock. The association of tephroite, spessartite, and rhodonite is one that has been recorded at a number of places, for example, Altarnun, Cornwall (Russell, 1946); Kaso Mine, Japan (Yosimura, 1939); and Broken Hill, Australia (personal observation). At Altarnun the mineral association probably represents a rhodochrosite deposit which has been subjected to contact metamorphism by a neighboring granite intrusion. The Kaso Mine deposit appears to be similar in origin. At Broken Hill the ore deposit containing the manganese silicates is located in an area of high-grade regional metamorphism, also characterized by almandine-cordierite-sillimanite gneisses.

It is tempting to speculate on the possibility of the tephroite at Clark Peninsula being associated with economic minerals. Most other occurrences of tephroite have been in ore deposits, as at Langban, Sweden (iron-manganese), Franklin, New Jersey (zinc), and Broken Hill, Australia (lead-zinc). Mr. Sullivan's observations of malachite staining in the country rock on Clark Peninsula may be significant in this respect. It is to be hoped that further geological information will be obtained from this region.

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A SAMPLE HOLDER FOR DIFFERENTIAL THERMAL ANALYSIS
OF FUSIBLE OR REACTIVE SAMPLESJ. L. FITCH AND B. G. HURD, *Magnolia Petroleum
Company, Dallas, Texas.*

The usual sample-thermocouple arrangement in differential thermal analysis (DTA) involves direct placement of the differential thermocouple within the samples. With this arrangement, fusion or sintering of samples makes sample changes difficult. Often, thermocouples are damaged beyond repair in attempting to remove them from the samples. Furthermore, chemical attack by various materials may alter the electrical characteristics of the thermocouples, in addition to shortening their lives. The use of noble metal couples reduces these difficulties, but does not eliminate them.

The use of external thermocouples facilitates sample changes and greatly prolongs thermocouple life. However, there have been objections to all previously suggested external thermocouple arrangements, and none have gained popular acceptance. Increased base line drift and loss of sensitivity and/or resolution have been the most serious difficulties experienced with the external couples. These problems are reviewed in recent books by Smothers and Chiang (1958) and Mackenzie (1957).

The sample holder-external thermocouple arrangement described here, with amplification of the response, provides good sensitivity without serious base line drift. The holders are thin-walled platinum microcrucibles, 12 mm. deep and about 12 mm. in diameter. They have a capacity of 1.3 cc. The bottom of each crucible is fitted with a closed-end center well 3 mm. in diameter and 6 mm. deep to accommodate the external thermocouple junction. The cylindrical well is made from platinum foil and is soldered to the crucible with a platinum alloy solder which melts at 1500 C.

The differential thermocouples used are platinum-platinum, 13 per cent rhodium made from B. & S. No. 28 gauge wire. The junctions are butt-welded. The thermocouples are housed in two-hole insulating spaghetti, about 2.5 mm. in diameter, and are mounted in an alundum cylinder used to seal one end of a tubular furnace. Quartz tubing, extending through the support block and along most of the length of the spaghetti, encloses the thermocouples to add strength.

The samples are supported in the furnace by the thermocouples themselves. When the sample holders are slipped over the thermocouples, the junctions are automatically positioned in the center of the sample cruci-

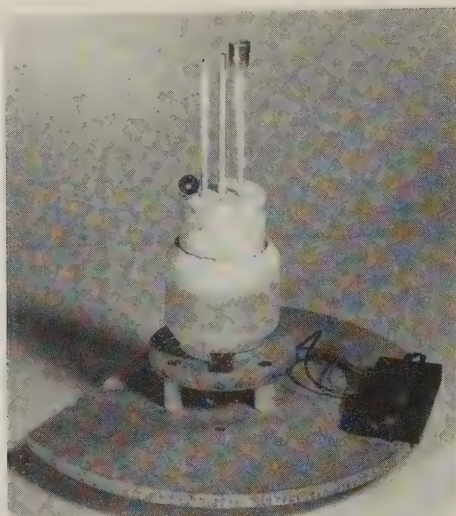


FIG. 1. DTA sample holder. External thermocouple assembly.

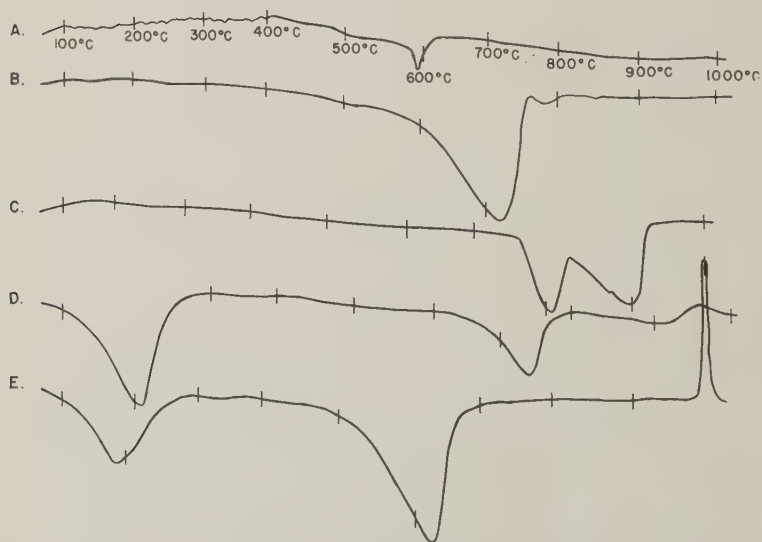


FIG. 2. DTA curves. (A) Quartz, Arkansas; amplification $\times 50$. (B) Magnesite, Luning, Nevada; $\times 20$. (C) Dolomite, Thornwood, New York; $\times 20$. (D) Montmorillonite, Osage, Wyoming; $\times 20$. (E) Halloysite, Eureka, Utah; $\times 20$.

bles. A thin coating of "Sauereisen" cement on the thermocouple junctions electrically insulates them from the sample holders.

The assembly, with one sample holder in place, is shown in Fig. 1. The other sample container, resting on top of the alundum block, is tilted to show the center thermocouple well. The center insulating tubing houses two thermocouples used to monitor the furnace temperature. They are connected to a recorder and temperature controller. The entire assembly, which may be exchanged for a conventional nickel block assembly, is supported by a collar which moves freely on a tubular stand (not shown). Exchange of the two types of sample holders is facilitated by quick-disconnect thermocouple lead connectors. By means of a clamp and locator-pin arrangement the samples are quickly and accurately positioned in the center of the furnace.

The quality of the DTA data obtained with the external thermocouple and platinum holders is illustrated by typical DTA curves shown in Fig. 2. The authors find the curves entirely satisfactory for most DTA purposes.

ACKNOWLEDGMENT

Permission of the Magnolia Petroleum Company to publish this note is gratefully acknowledged.

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SPONTANEOUS OXIDATION OF A SAMPLE OF POWDERED SIDERITE*

WALDEMAR T. SCHALLER AND ANGELINA C. VLISIDIS,
U. S. Geological Survey, Washington, D. C.

In checking over the analyzed samples left by the late Dr. Roger C. Wells, former geochemist, U. S. Geological Survey, it was noted that a powdered sample from near Linden, Texas, labeled siderite, had a marked reddish color, suggesting considerable oxidation of the ferrous

* Publication authorized by the Director, U. S. Geological Survey.

iron. A determination of the ferrous iron content, made by J. J. Fahey in 1944, yielded the unexpected low result of only some six per cent FeO, a very considerable decrease from the original figure of 59 per cent determined by Wells in 1915. About 90 per cent of the original FeO content had spontaneously oxidized to Fe_2O_3 within a period of 29 years. Since 1944 the oxidation has continued and now (May 1958) the percentage of FeO is less than one per cent. The color of the sample in May 1958 is very close to Ridgway's "Mahogany Red," Plate II, 7. R-O. k, and its smear is very similar to the adjoining "Burnt Sienna."

The occurrence of this siderite iron ore from the vicinity of Linden, Cass County, Texas, is given by Burchard (1916, p. 79-86). The analysis (Wells, 1937, p. 95-96) was made on an "average sample cleaned carbonate ore." As was customary in those days, the sample was powdered in a mechanically driven agate mortar, probably running for the better part of a day. Wells' analysis (FeO 59.42, MnO 0.13, CaO 0.16, CO_2 36.54, SiO_2 1.10, Fe_2O_3 0.40, Al_2O_3 1.02, P_2O_5 0.11, S 0.05, H_2O^- 0.15, H_2O^+ 1.05, TiO_2 0.04, total 100.17) indicates an unusually pure siderite, the FeO content of the siderite being raised to 61.74 per cent (calculated percentage of FeO for FeCO_3 being 62.01) after deducting nearly four per cent of impurities. These figures indicate that very little if any FeO was oxidized during the grinding.

The powdered sample was preserved in a glass tube closed with a cork stopper and lay in Wells' laboratory drawer undisturbed from 1915 to 1944 except for the transfer involved in moving into a new building in 1917. About 1950 the sample was transferred to a screw-top glass jar.

After Fahey's determination of FeO in 1944, no additional determination was made for a period of 10 years. In 1954, a determination of FeO gave a value less than half the amount found by Fahey, and it became evident that the ferrous iron content of the powdered sample was being spontaneously reduced as a function of time, and more frequent determinations were then run.

The results so far obtained, for decreasing percentages of FeO in a sample of powdered siderite as a function of time are as follows:

Date	Time in Years	Analyst	Percentage FeO
March 1915	—	R. C. Wells	59.42
May 1944	29	J. J. Fahey	6.23
May 1954	39	A. C. Vlisidis	2.38
May 1956	41	A. C. Vlisidis	1.70
June 1957	42	A. C. Vlisidis	1.20
May 1958	43	A. C. Vlisidis	0.74

In September 1955, Fred A. Hildebrand (U. S. Geological Survey, written communication) reported, on the basis of an x-ray pattern (film no. 8675), that the sample was "... hematite with small to moderate amounts of goethite and siderite and trace of quartz. The hematite appears to be poorly crystalline and has a unit cell somewhat larger than normal hematite." The trace of quartz is 1.10 per cent of SiO_2 by Wells' analysis.

Plotting the percentages of FeO shown above against time (years) and extending a straight line through the last five points, suggests that in early 1960, about 45 years after Wells' analysis, all the ferrous iron will have spontaneously oxidized to ferric iron.

These results lead to speculations on the history of natural deposits of ferric iron oxides. How many of these were at one time ferrous carbonates? These speculations we leave to the geologists. Burchard (1916, p. 76) noted the partial oxidation of nodular masses of siderite to iron oxide on Bowie Hill, Cass County, Texas, stating, "The iron carbonate is in general partly altered to limonite or to reddish hydrated oxides of iron, which form a scale or crust of varying thickness around the carbonate nucleus and along cracks which intersect the masses." A massive sample of siderite perhaps could be protected from further oxidation by a surface layer of ferric oxide whereas no such crust of oxidized iron would form in a powdered sample.

What was it that caused the spontaneous, almost complete oxidation of this sample in approximately 43 years? Specimens of siderite in collections remain unoxidized for much longer time. Was there a latent after reaction due to the mechanical and thermal effects produced by, induced by, or accompanying the grinding of the sample?

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A QUANTITATIVE CORRECTION FOR THE HOLMES EFFECT

JOHN W. CAHN, *General Electric Research Laboratory, Schenectady, N. Y.*

The Holmes effect (1) arises in modal analysis because it is impossible to confine the observation to a single plane in a transparent specimen.

Thus when looking at a thin slice there is a tendency to measure the particle's projected area instead of its area of intersection with the surface of the slice. This leads to an over-estimate of the area fraction and also of the volume fraction which is derived from it. The correction is especially serious for small grain diameters or cross sectional areas.

In a recent paper on quantitative transmission metallography, Cahn and Nutting (2) developed a set of expressions which relate quantities measured on the projected plane to parameters in the metal specimen. One of these relations can be used for an accurate quantitative correction for the Holmes effect.

The expected projected area fraction f_A of a phase, for a slice of thickness t , is given by (2)

$$f_A = f_V + \frac{1}{4}St \quad (1)$$

where f_V is the volume fraction of this phase and S is the total surface area per unit volume. This equation holds true for convex particles of any shape although it ignores overlap of two or more particles on the plane of projection. This overlap is only a problem when the slice thickness is many times the grain diameter and need not concern us here. A first order correction was developed by Cahn and Nutting.

The derivation of the equation is quite straightforward. Consider the elements of grain surface in the interior of the slice. If the slice is obtained by a random section, these surfaces on the average will be so inclined that their projected area is half of the actual area (3). However the surfaces created by the intersection of the grains with the plane of section are always normal to the light beam, and thus their projected area is equal to their actual area. If the particles are convex then every element of projected area results from two surfaces in the specimen.

From the Delesse relation (1) the area fraction on the plane of section is expected to be equal to f_V . The surface area in the interior of the slice is expected to be St . Thus the equation is proven.

In order to obtain a useful correction for the Holmes effect S must be evaluated. This can be done in a quite straightforward manner based on relations derived by Smith and Guttman (4). They give for S either

$$S = \frac{4l}{\pi A}$$

or

$$S = \frac{2N}{L}$$

where l is the total particle perimeter observed on plane of section and A

is the section area, or N is the number of intercepts of grid lines with the particles or grain surfaces and L is the total length of grid lines. Thus

$$f_v = f_A - \frac{Nt}{2L}$$

or

$$f_v = f_A - \frac{lt}{\pi A}. \quad (2)$$

It should be pointed out that in the measurement of S there will again exist an effect analogous to the Holmes effect in that the perimeter, or the number of intersections will appear to be larger than they actually are. This is a second order correction, however, which may also be easily estimated.

ACKNOWLEDGMENT

The author wishes to thank J. E. Hilliard for calling his attention to the Holmes effect.

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GRAPHICAL REPRESENTATION OF AMPHIBOLE COMPOSITIONS*

J. V. SMITH, *The Pennsylvania State University,
University Park, Pennsylvania.*

Compositional variations of amphiboles are complex and difficult to memorize. As data are generally more easy to assimilate in a visual than in a mathematical form, a graphical representation is desirable.

Amphiboles fall readily into two groups,

anthophyllite and cummingtonite formula $A_2B_5C_8O_{22}D_2$

where

A is largely Mg, Fe²
B is Mg, Fe², Al, Fe³ etc.
C is Si, Al
D is OH, F, Cl.

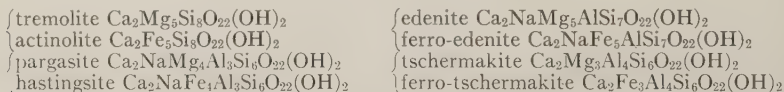
* Contribution No. 58-21, College of Mineral Industries, Pennsylvania State University. University Park, Pennsylvania.

calcium-sodium amphiboles formula E_2 to $3B_5C_8O_{22}D_2$

where E is Ca, Na, K etc.

In the first group the two important substitutions are Mg for Fe^2 and 2Al for Si + (Mg, Fe^2). Consequently this group of amphiboles can be graphically displayed on a square diagram with corners $Mg_7Si_8O_{22}(OH)_2$, $Fe_7Si_8O_{22}(OH)_2$, $Mg_5Al_2Si_6Al_2O_{22}(OH)_2$, $Fe_5Al_2Si_6Al_2O_{22}(OH)_2$ as is already well known (Winchell and Winchell, 1951).

In the second group, compositional variations are more complex, for there is interaction between ions of four different valencies. Sundius (1946) has determined a complete list of end members and his results and nomenclature will be adopted here. Hallimond (1943) has devised a graphical representation for calcic amphiboles with Ca = 2. The diagram consists of a quadrilateral (Fig. 1) with apices



In this diagram Mg and Fe are included together; by erecting a vertical ordinate on this base the variation between the Mg and Fe members can be shown, as, for example, in Winchell and Winchell (1951).

If, instead, a vertical ordinate is used that expresses the variation $CaAl \rightarrow NaSi$, the relationship between calcium and sodium amphiboles

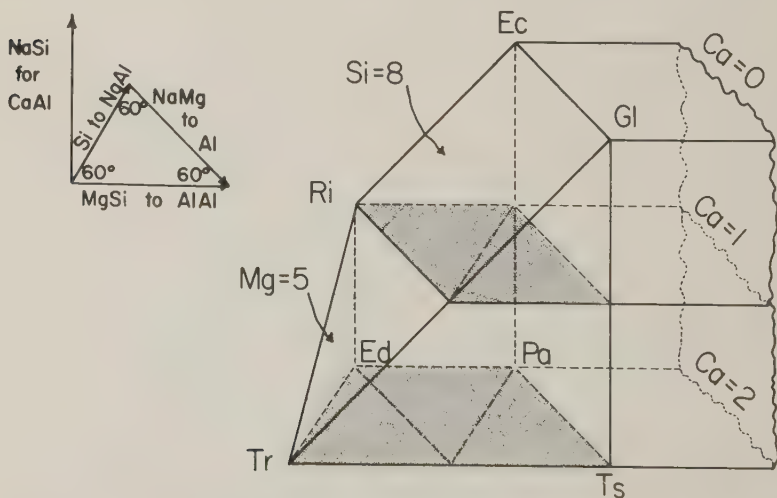


FIG. 1. Three-dimensional diagram showing the compositional variation in amphiboles. Tr, tremolite; Ts, tschermakite; Ed, edenite; Pa, pargasite; Ri, richterite; Ec, eckermanite; Gl, glaucophane. The inset diagram shows the atomic substitutions that form the basis of the large diagram.

can be readily seen (Fig. 1). When all the Ca has been exchanged for Na, pargasite (hastingsite) and tschermakite (ferro-tschermakite) are transformed into eckermannite $\text{Na}_3\text{Mg}_4\text{AlSi}_5\text{O}_{22}(\text{OH})_2$ (arfvedsonite $\text{Na}_3\text{Fe}_4\text{AlSi}_5\text{O}_{22}(\text{OH})_2$) and glaucophane $\text{Na}_2\text{Mg}_3\text{Al}_2\text{Si}_8\text{O}_{22}(\text{OH})_2$ (riebeckite $\text{Na}_2\text{Fe}_3\text{Al}_2\text{Si}_8\text{O}_{22}(\text{OH})_2$). When only one Ca has been exchanged edenite (ferro-edenite) is turned into richterite $\text{CaNa}_2\text{Mg}_5\text{Si}_8\text{O}_{22}(\text{OH})_2$ (ferro-richterite $\text{CaNa}_2\text{Fe}_5\text{Si}_8\text{O}_{22}(\text{OH})_2$).

The possible compositional variations of calcium-sodium amphiboles can be deduced from a crystal-chemical study of the amphibole structure, and readily visualized from the diagram. The upper limit of the diagram is simply fixed by the requirement that Ca does not fall below zero. Only rarely in published analyses is Ca greater than 2, and perhaps these exceptions are caused by analytical error. This fixes the lower limit of the diagram. The third limiting factor is that Si cannot be greater than 8 as only eight sites of tetrahedral coordination exist in the amphibole structure. In the diagram the plane with $\text{Si}=8$ passes through eckermannite, glaucophane, tremolite and richterite. All compositions to the left of this would have Si greater than 8 and therefore are forbidden. The fourth factor is that $\text{Ca}+\text{Na}$ cannot be greater than 3, the number of large holes in the structure. Thus all compositions behind the plane defined by edenite, richterite, eckermannite and pargasite are forbidden. Fifthly, the sum of $\text{Ca}+\text{Na}$ does not fall below 2 except in a few analyses, the structural reason apparently being that the large holes occupied in anthophyllite must be completely filled in all amphiboles. All compositions in front of the plane tremolite, glaucophane and tschermakite are therefore prohibited. The $\text{Mg}+\text{Fe}^{2+}$ content provides the sixth criterion, for only 5 sites in the structure are available for these ions. Consequently all compositions to the left of the plane tremolite, edenite and richterite are forbidden. The final criterion is a practical one. Compositions to the right of the plane defined by pargasite, tschermakite, glaucophane and eckermannite rarely occur, which means that the substitution of Al in the amphibole structure is limited to considerably less than the theoretical maximum. Thus the compositional variation of hornblendes is limited to a wedge-shaped region with one corner cut off.

Difficulties arise when actual analyses are plotted on the diagram. For each of the four valence states, planes can be erected which express the amounts of the ions. Ideally the four planes should meet in a point but in practice discrepancies occur, some of them very large. The reasons for these have not yet been fully evaluated, although four can be readily suggested; error in the chemical analysis together with impurity in the analyzed material, occurrence of Mg and Fe in the large holes normally occupied by Na and Ca (equivalent to solid solution between a horn-

blende and the anthophyllite-cummingtonite series); replacement of OH^- by O^{--} (the oxy-hornblende reaction) or the presence of an atomic substitution not considered in Sundius's scheme. The relative significance of these possibilities together with a way of recognizing them will be considered in a later publication. The diagram is presented at this time because of the hope that it will prove of help (especially to students) in visualizing the broad compositional variations of amphiboles.

The assistance obtained from a National Science Foundation grant is gratefully acknowledged.

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THE NEAR INFRARED SPECTRUM OF BERYL

KENNETH A. WICKERSHEIM AND ROBERT A. BUCHANAN, *University of California, Los Angeles, California.*

Several references to the Raman and infrared absorption bands of beryl in the 2.7 micron (3700 cm^{-1}) region have appeared in the literature during the past twenty-six years. Nisi (1) observed a Raman band at 3607 cm^{-1} using as a sample a clear, light green, hexagonal prism of beryl. No attempt was made to account for this frequency. Later Matossi and Bronder (2) reported the infrared spectrum of a sample of aquamarine taken with an instrument of low resolving power. Two strong bands were observed near 3600 and 3700 cm^{-1} . In addition weaker clusters of bands were observed in the 5000 and 7000 cm^{-1} regions. Matossi and Bronder felt that these bands were too intense to be due to a water impurity in the beryl, arguing that a 1 per cent impurity would be possible, but that the intensity of the bands would indicate a 5 to 10 per cent impurity. They suggested that the bands could be due to a hydroxyl impurity and to multiple combinations of silicate frequencies.

Lyon and Kinsey (3) studied a narrow region of the spectrum in the vicinity of 3700 cm^{-1} under higher resolution using a grating spectrometer and a sample of beryl 0.2 mm. thick. They found two intense bands whose centers were estimated to be 3598 and 3690 cm^{-1} plus a weak shoulder on the high frequency side of the second band. They identified these bands with envelopes of certain rotation branches of the sym-

metric and antisymmetric stretching frequencies of the water molecule and accounted for the freedom required for rotation by means of an interesting hypothesis. They suggested that the water occurred as a monomeric impurity in the large open channels of the beryl crystal.

Eitel (4) also refers to the possible occurrence in these channels of molecular impurities. Among the impurities he names water and several alkali hydroxides.

Wilmarth, Warner, Hosler and Cameron (5) state that "alkalis and alkaline earths, principally soda, constitute as much as 4 per cent of many beryls, and constitutional water up to 2.7 per cent." These authors suggest the possibility of the occurrence of hydroxyl ions substituted for oxygens in the silicate rings of the crystal itself.

The articles referred to above suggest three possible explanations for the near infrared absorption bands of beryl. The bands may be (1) due to water molecules either free or weakly bound in the channels, (2) due to alkali or alkaline earth hydroxide molecules in the channels, or (3) due to hydroxyl groups incorporated into the beryl structure through substitution for oxygen.

Because of their frequencies there seems to be little doubt but that the fundamentals observed in the 3700 cm^{-1} region arise from OH stretching vibrations associated with hydroxyl or water impurities. The work to be reported here represents an attempt to see what further information could be obtained from a purely spectroscopic study of beryl. Several different samples of beryl were studied to determine whether the impurities varied from sample to sample. Polarized radiation was employed to determine the degree of orientation of the impurities relative to the crystallographic axes. And all spectra were taken under high resolution over a much wider region than that observed by Lyon and Kinsey.

The grating spectrometer used in this study was the same one used by Lyon and Kinsey (3) but with its range extended and modified to operate with an AC detection system. A selenium film polarizer was mounted just in front of the sample at the exit slit of the instrument. The spectral slit width of the spectrometer varied from 3 cm^{-1} at 3 microns to about 10 cm^{-1} at 1 micron. All band centers quoted in this article are believed to be accurate to within 3 cm^{-1} .

The beryl samples were selected from gem lots provided by the O'Brien Lapidary Equipment Company of Los Angeles. These lots were labeled only as to national origin. Sample 1 was a clear, medium blue, hexagonal prism 15 mm. long and 5 mm. in diameter from Madagascar. Samples 2, 3 and 4 were all rough fragments from Brazil and were respectively pale blue, medium pink and colorless. All samples were clear except Sample 2 which had fibrous inclusions.

Sample plates were cut from each specimen of beryl using standard faceting techniques and equipment. These plates were thinned in steps with absorption spectra being taken at each step.

The polarized absorption spectra obtained from the four pieces of beryl are shown in Fig. 1. Certain facts are immediately evident from the spectra. The frequencies of absorption are not the same in all samples. For example in the spectrum of Sample 3 no absorption bands agree with the bands of any other sample. On the other hand, where identical absorption bands do occur in different samples their intensities sometimes differ strikingly for similar sample thicknesses. And the relative intensities of the bands within a spectrum sometimes differ from sample to sample. These effects indicate that more than one type of molecular impurity can occur in beryl and that the concentration of each type of impurity can vary considerably.

Another obvious fact is the high degree of orientation of the absorbing entities relative to the *c* axis of the crystal as indicated by the strong dichroism of the bands. This would seem to rule out the occurrence of free molecular rotation within the crystal.

Let us consider for the moment the spectra of Samples 1, 2 and 4, since these show distinct similarities to one another while the spectrum of Sample 3 is completely dissimilar. The three strongest bands observed in these samples are centered at 3598 and 3700 cm^{-1} (both polarized parallel to the *c* axis) and at 3663 cm^{-1} (polarized perpendicular to the *c* axis). The first two bands are apparently those observed by Matossi and Bronder and by Lyon and Kinsey. The two stretching fundamentals of the water molecule should be polarized at right angles to one another. Thus the first two bands mentioned above cannot both be associated with the two water stretching vibrations. As further evidence for this statement we note that in Sample 1 the band at 3700 cm^{-1} is distinctly stronger than the band at 3598 cm^{-1} while in Sample 4 the band at 3598 cm^{-1} is the stronger of the two. It would appear that these two bands arise from entirely different impurities.

We might however relate the antisymmetric stretching frequency of the water molecule to the 3700 cm^{-1} band and the symmetric stretching frequency to the 3663 cm^{-1} band. First and second overtones of these two fundamentals appear in the 7100 and 10,400 cm^{-1} regions. In addition strong combinations of these fundamentals and overtones with a frequency of 1540 to 1600 cm^{-1} (or a multiple thereof) occur, giving rise to bands in the 5300, 6800 and 8700 cm^{-1} regions. The water molecule possesses a third (bending) mode of vibration, the fundamental frequency of which falls in the vicinity of 1600 cm^{-1} . The appearance of strong combinations involving such a frequency is probably the best evidence that it is water which gives rise to the 3663 and 3700 cm^{-1} fundamentals.

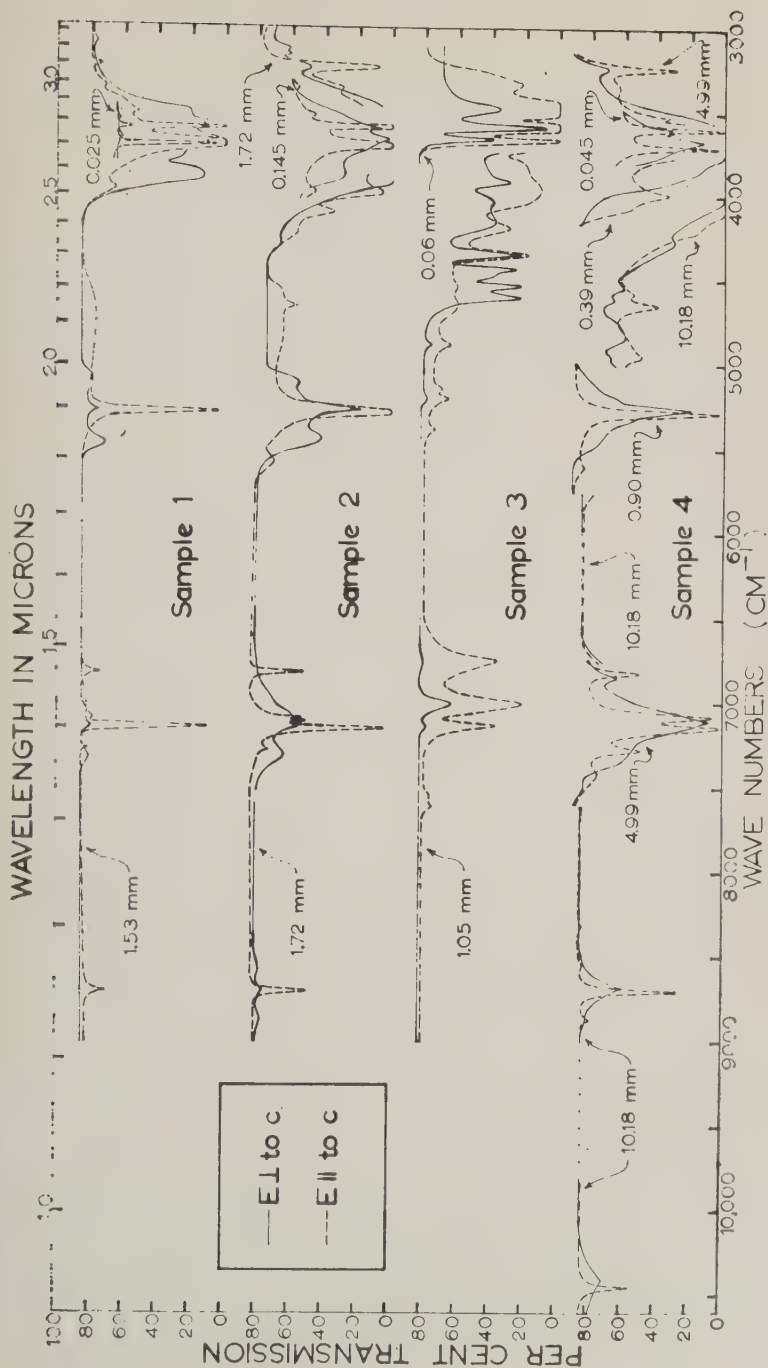


FIG. 1. Absorption spectra of four different samples of beryl.

The high frequency of these fundamentals and the narrowness of the bands would seem to indicate that the water molecules are monomeric and at most only weakly bound to the lattice. On the other hand, the strong polarization of the bands would imply that the water molecules are strongly oriented with their H—H distances parallel to the c axis of the crystal. It seems most likely then that the water molecules occur singly in the channels with the above-mentioned orientation.

The fundamental at 3598 cm^{-1} is also polarized parallel to the c axis of the crystal. This band comes the nearest of any of the strong fundamentals to agreeing with the Raman band observed by Nisi at 3607 cm^{-1} . The strong orientation of this band, its high frequency and narrowness and its apparent independence of other bands in this region combine to suggest that it belongs to an alkali or alkaline earth hydroxide molecule lying lengthwise in the beryl channels.

Numerous weaker bands are found in the regions immediately adjoining the strong fundamentals mentioned above. Some of these bands appear to be low frequency combinations with the fundamentals. For example at 170 cm^{-1} the frequency appears to combine to produce both summation and difference bands with the 3700 cm^{-1} fundamental and with its overtones and combinations. This 170 cm^{-1} frequency could correspond to a weakly hindered rotation of the water molecule.

On the other hand many of the weaker bands seem to be unrelated to any of the fundamentals previously mentioned. These bands may be fundamentals in their own right arising from weaker impurities, or they may be combinations and overtones of lower, unobserved vibration frequencies.

The spectrum of Sample 3 is very different from the spectra of the other samples. There are three fundamentals, all imperfectly polarized but strongest parallel to the c axis, with centers at 3479 , 3583 and 3653 cm^{-1} . First overtones of these fundamentals are found at 6745 , 6997 and 7138 cm^{-1} . A complex set of weaker absorption bands appears on the high frequency side of the three fundamentals. The fundamentals of Sample 3 show less dichroism and are somewhat broader than the fundamentals of Samples 1, 2 and 4. Thus if any of the samples of beryl contain bound hydroxyl groups substituting for oxygen, Sample 3 is the most likely prospect.

It is clear that this study has not solved the entire problem. In order to relate the spectra to specific impurities it would be best to correlate the spectroscopic study with a program of controlled synthesis of beryl. The idea of such a synthesis with controlled impurities is appealing from the standpoint of the physicist or chemist since it contains the possibility of isolating and orienting for spectroscopic study a variety of small

molecules in the beryl channels. On the other hand, assuming that a detailed identification of absorption bands can eventually be made, the infrared spectrum of beryl should be useful to the mineralogist in the identification of impurities and in the determination of their concentrations.

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THE SYNTHESIS OF UVAROVITE

S. GELLER AND C. E. MILLER, *Bell Telephone Laboratories, Incorporated, Murray Hill, New Jersey.*

As part of another investigation, we recently attempted the preparation of the garnet $\text{Ca}_3\text{Cr}_2(\text{SiO}_4)_3$. The direct synthesis of this garnet has been reported by Hummel (1) whose results we have corroborated only in part.

The largest amount of garnet phase was produced by heating the pressed mixture of reactants ($3\text{CaO} + \text{Cr}_2\text{O}_3 + 3\text{SiO}_2$) at 1400°C . for 2 hrs. The extra phases present were $\alpha\text{-CaSiO}_3$ and Cr_2O_3 . Three samples heated at 1200°C . for 47, 125 and 192 hrs. gave only very small amounts of garnet phase and large amounts of the $\alpha\text{-CaSiO}_3$ and Cr_2O_3 .

The powder data for the garnet phase are given in Table 1. The lattice constant of the synthetic uvarovite is $12.00 \pm 0.02 \text{ \AA}$.

Menzer (2) has reported a lattice constant of $11.974 \pm 0.003 \text{ \AA}$ for an uvarovite specimen from Sysmä, Finland. The chemical analysis of this sample indicated the presence of 1.93% Al_2O_3 , 0.41% Fe_2O_3 and 0.50% MgO impurities. A sample from Bissersk in the Urals reported by Menzer had a lattice constant of $11.969 \pm 0.013 \text{ \AA}$ but larger amounts of Al_2O_3 (5.8%) and MgO ($\sim 1.3\%$). This sample also had larger amounts of iron, probably 2.0% in terms of Fe_2O_3 , than did the sample from Finland.

We have obtained two samples of natural uvarovite of which the lattice constants have been measured but no chemical analyses made. One, obtained from Ward's Natural Science Establishment, originally found in Orford, Quebec, was crystallized on tremolite. The crystals were

TABLE 1. POWDER DATA FOR SYNTHETIC UVAROVITE
(CrK α RADIATION)

<i>hkl</i>	<i>d_{obs}</i>	<i>d_{calc}</i>	I
321	3.21	3.21	w
400	3.00	3.00	m-s
420	2.68	2.68	s
332	2.56	2.56	vw
422	2.45	2.45	m
431	2.35	2.35	w
521	2.19	2.19	w
611, 532	1.95	1.95	w-m
620	1.90	1.90	vw
444	1.731	1.732	vw
640	1.663	1.664	m
642	1.602	1.604	m-s
800	1.499	1.500	w
840	1.341	1.342	w-m
842	1.309	1.309	m
565, 921, 761	1.294	1.294	w
664	1.279	1.279	w-m

very small and included an impurity phase. The other sample* from Outokumpu, Finland, was obtained from Dr. G. Switzer of the Smithsonian Institute. Both samples have the same lattice constant, 11.931 ± 0.005 Å, within experimental error. This value is significantly smaller than that of the synthetic; presumably there are substantial amounts of Al³⁺ and possibly Fe²⁺ and/or Mg²⁺ ions in these samples.

Recently, Gillery (3) reported the lattice constant 11.87 Å for an uvarovite specimen from Orford, Quebec. However, that specimen contained 36.7% SiO₂, 33.2% CaO, 17.5% Al₂O₃, 6.2% Cr₂O₃, 5.0% FeO, 0.81% MgO. Thus, for Gillery's sample, a lattice constant much closer to that of grossularite, Ca₃Al₂(SiO₄)₃ should be expected; for a pure synthetic grossularite, the lattice constant is 11.851 Å (4)

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* Smithsonian sample designation No. 106829.

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FURTHER STUDIES ON $6\text{CaO} \cdot 3\text{SiO}_2 \cdot \text{H}_2\text{O}^*$ LESLEY DENT GLASSER AND DELLA M. ROY, *The Pennsylvania State University, University Park, Pennsylvania*

INTRODUCTION

In the course of studies in the system $\text{CaO-SiO}_2\text{-H}_2\text{O}$ (1), several new calcium silicate hydrates were prepared. One of these, C_2SH_x ,[†] termed "Phase Y" by Roy, was found to be in equilibrium with $\alpha'\text{-C}_2\text{S}$ and water above 800°C . at about 1000 atm. water vapor pressure. Determination of weight loss on heating to 1250°C . gave the formula $\text{C}_6\text{S}_3\text{H}$.

Small single crystals have been obtained, and the unit cell determined. The power data have been indexed, and in addition dehydration and D.T.A. curves have been obtained.

UNIT CELL

The compound crystallizes in small prisms. Rotation and oscillation photographs about all axes, together with $0kl$, $1kl$, $2kl$, $h0l$, and $hk0$ Weissenbergs were taken. They showed the unit cell to be triclinic; the cell constants, refined by comparison with the powder data, are given in Table 1. Both real and reciprocal spacings are quoted, since the latter are often the more useful. The prism axis is a . Evidence was found of twinning on (010).

The powder data were indexed by direct comparison of rotation and

TABLE 1. UNIT CELL PARAMETERS FOR $6\text{CaO} \cdot 3\text{SiO}_2 \cdot \text{H}_2\text{O}$

Real cell		Reciprocal cell ($\lambda = 1.542\text{ \AA}$)	
a	6.84 \AA (all $\pm .02$)	a^*	.2297 r.u. (all $\pm .0005$)
b	6.94	b^*	.2244
c	12.89	c^*	.1206
α	$90^\circ 45'$ (all $\pm 10'$)	α^*	$88^\circ 10'$ (all $\pm 3'$)
β	$97^\circ 22'$	β^*	$82^\circ 27'$
γ	$98^\circ 16'$	γ^*	$81^\circ 33'$

$$V = 601\text{ \AA}^3 \quad G = 2.94 \quad Z = 1.99.$$

* Contribution No. 58-34. The Pennsylvania State University, College of Mineral Industries, University Park, Pennsylvania.

[†] Standard cement chemical nomenclature ($\text{C} = \text{CaO}$, $\text{S} = \text{SiO}_2$, $\text{H} = \text{H}_2\text{O}$) will be used throughout.

powder photographs taken on the same 6cm. diameter camera, and the indices assigned were checked against a list of the strongest intensities found on the Weissenberg photographs. Once the indices of most lines had been fixed by this method, the cell parameters were refined to give the best agreement with accurate d values; these were obtained using a Norelco wide range diffractometer with filtered copper radiation and 1° slit. The instrument had been calibrated against a silicon standard.

It was considered impracticable to carry the indexing below about $d = 2 \text{ \AA}$, since at such small spacings it is difficult to obtain unambiguous indices. Down to this point the indexing is satisfactory, and the agreement between observed and calculated d values is good.

D.T.A. CURVES

A differential thermal analysis was made on two samples dried at 110° C. ; quartz was used as reference standard and the rate of heating was $5^\circ/\text{min.}$ In each case, one large endothermic peak was found, at 700° C.

DEHYDRATION CURVES

Dehydration curves were determined by heating the sample in a horizontal furnace regulated by a temperature controller; the temperature was measured by a chromel-alumel thermocouple placed immediately above the sample. Nitrogen was passed through tubes freshly packed with soda-lime and silica gel and then through the furnace to provide an inert atmosphere. About 100 mg. of sample was contained in a 300 mg. Pt microboat. This was heated to successively higher temperatures and weighed, at intervals of 12 hours, to essentially constant weight (within about 0.1 mg.). Extra sample was placed in the furnace, to permit portions to be removed and examined at various stages.

Runs were made on two samples (Fig. 1), both of which had previously been dried at 100° C. In each case there was a loss at about 550° C. and a second, larger, loss at about 650° C. After complete dehydration, the product was $\beta\text{-C}_2\text{S}$. The loss at 650° was the same for both samples, and agrees with the water content as originally determined by Roy. It presumably corresponds with the 700° D.T.A. peak and is not inconsistent with the data of Roy (1) for the equilibrium hydrothermal decomposition of Phase Y. The loss at 550° was much greater in sample A than in sample B; a portion of A was withdrawn at the point P in Fig. 1, and its diffraction trace showed that it was still essentially Phase Y. This confirms Roy's formula, but leaves the earlier loss unexplained. Examination of the diffraction patterns of the starting materials showed that sample A had two lines with pronounced intensity differences as compared with the

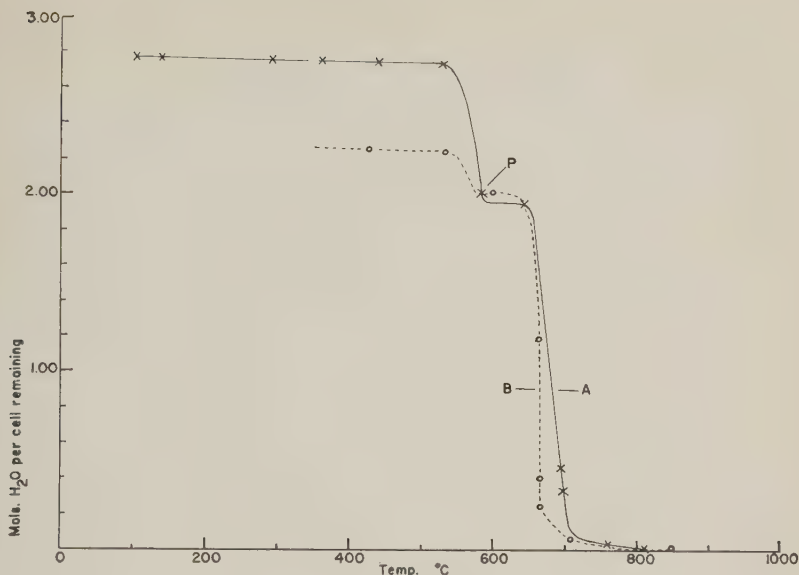


FIG. 1. Dehydration curves for samples A and B (see text). Principal weight-loss at about 650° corresponds to dehydration of Phase Y. Sample removed at P is essentially unchanged.

standard pattern for Phase Y; these lines are indicated with an asterisk in Table 2. In the sample withdrawn at point P, the intensity of these lines was normal, and a new line had appeared at about 3.12 \AA . Sample B before heating also showed this intensity difference, but to a much smaller degree. The correlation between intensity variation and water loss at 550° suggested that a small amount of some other phase might be present. Re-examination of diffraction traces from the earlier work revealed that slight intensity differences had occasionally shown up before, and that once or twice the line at 3.12 \AA had been found. In all cases the effects had been very small, and, since they had not been consistent, had been regarded as spurious. The extra lines did not appear to correspond to any of the known calcium silicates, hydrous or anhydrous, or to lime or calcite.

Optical examination of the initial material showed that a small amount of a fibrous substance was present. Some of this was sorted out by hand, and a powder photograph was taken. The pattern obtained, which showed strong preferred orientation effects, was similar to that of hillebrandite. This is consistent with the observed dehydration temperature, 550° . The two lines noted above in the diffractometer pattern might conceivably correspond to those noted at 3.33 and 3.02 \AA in natural

TABLE 2. POWDER DATA AND INDICES FOR $6\text{CaO} \cdot 3\text{SiO}_2 \cdot \text{H}_2\text{O}$

$d_{\text{obs.}}$	Int.	$d_{\text{calc.}}$	hkl	d_{obs}	Int.	$d_{\text{calc.}}$	hkl
6.90	25	$\begin{cases} 6.87 \\ 6.71 \end{cases}$	$\begin{cases} 010 \\ 100 \end{cases}$	2.736	10	2.747	2 $\bar{1}2$
				2.718	20	$\begin{cases} 2.717 \\ 2.710 \end{cases}$	$\begin{cases} 02\bar{3} \\ 2\bar{1}\bar{3} \end{cases}$
4.61	10	4.61	012			2.633	023
				2.634	15	$\begin{cases} 2.630 \\ 2.632 \end{cases}$	$\begin{cases} 12\bar{3} \\ 114 \end{cases}$
3.435	70	3.435	020			2.599	220
*3.349	25	3.356	$\begin{cases} 20\bar{1} \\ 200 \end{cases}$	2.594	2	2.556	005
				2.556	25	2.483	20 $\bar{4}$
3.293	20	3.291	021	2.488	2	2.475	22 $\bar{2}$
3.258	25	3.258	120	2.473	2	$\begin{cases} 2.291 \\ 2.290 \end{cases}$	$\begin{cases} 105 \\ 030 \end{cases}$
3.209	5	3.207	2 $\bar{1}0$	2.290	100	2.275	214
*3.067	50	3.067	02 $\bar{2}$			$\begin{cases} 2.272 \\ 2.255 \end{cases}$	$\begin{cases} 130 \\ 30\bar{1} \end{cases}$
				2.270	5	2.159	301
2.986	40	2.986	022	2.243	2	$\begin{cases} 2.134 \\ 2.135 \end{cases}$	$\begin{cases} 032 \\ 203 \end{cases}$
2.890	25	2.891	120			2.823	202
2.860	15	$\begin{cases} 2.863 \\ 2.864 \end{cases}$	$\begin{cases} 21\bar{1} \\ 122 \end{cases}$	2.135	5		
2.823	35	2.823	$\begin{cases} 20\bar{3} \\ 202 \end{cases}$				

hillebrandite (2), although it has been pointed out (2) that the synthetic material usually shows shorter, rather than longer, spacings. Preferred orientation could account for the apparent absence of the rest of the pattern in the diffractometer trace. On the whole, the evidence is not too satisfactory on this point, but it seems fair to assume that the extra phase is a hillebrandite-like material. If the extra phase is assumed to have a water content similar to that of hillebrandite, recalculation of the water content of Phase Y raises it to 2.1–2.2 mols. per cell, which does not materially alter the conclusion regarding the formula.

CELL CONTENTS

It was felt that it would be hopeless to attempt to determine the density of such tiny crystals by conventional methods. The density was therefore calculated from the refractive indices of the crystals, using the adaptation of the Lorentz-Lorenz equation derived by Howison and Taylor (3), which usually gives values correct to within $\pm 2\%$. This gives a value of 2.94; on this basis $Z=1.99$, and the cell contents are $12\text{CaO} \cdot 6\text{SiO}_2 \cdot 2\text{H}_2\text{O}$.

The conditions of formation and dehydration suggest that the water may be present as hydroxyl groups not attached to silicon. Gard and

Taylor (4) point out that calcium silicates containing water as molecules, or as hydroxyl attached to silicon, such as afwillite (5) or tobermorite (6), tend to be formed hydrothermally, and to dehydrate, at much lower temperatures. The temperature of dehydration of Phase Y is much more nearly comparable to those of xonotlite (7) and foshagite (4) which are considered to contain water as hydroxyl groups not attached to silicon. For xonotlite, this has been established by a structure determination (8). If Phase Y really does contain its water as hydroxyl not attached to silicon, the constitution (for the unit cell contents) is $\text{Ca}_{12}\text{Si}_6\text{O}_{22}(\text{OH})_4$, which implies the presence of Si_2O_7 groups, together with either SiO_4 groups or oxide ions. Such a mixture of groupings has been reported for epidote and zoisite (9), and it might well account for the awkwardly shaped unit cell. On the other hand, the prism axis, a , (6.84 Å) is similar in length to b (6.76 Å) for β - C_2S (or c , the prism axis of α' - C_2S). Together with the high refractive indices, this suggests a possible similarity in structure. However, examination of the pseudomorphs formed by dehydrating the original material gave no evidence of preservation of order, implying that there is no simple relationship. Without further work, it is not possible to make even a guess at the structure.

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POLYMORPHISM IN BARIUM DISILICATE

ROBERT S. ROTH AND ERNEST M. LEVIN, *National Bureau of Standards, Washington, D. C.*

The question of polymorphism in barium disilicate has recently been discussed in this journal by R. M. Douglass (1958). Although Rogers (1932) postulated dimorphism of BaSi_2O_5 in order to reconcile his optical data with those of Bowen (1918) and Escola (1922), Douglass found no evidence of polymorphism in his own experiments. Douglass compared his indexed x -ray diffraction powder pattern of the mineral sanbornite with the pattern published by Austin (1947) for $\text{BaO} \cdot 2\text{SiO}_2$. He correlated the (112/004) doublet of sanbornite occurring at 3.422 \AA with the 3.55 \AA peak in Austin's pattern. However, if a calculation of Austin's pattern is made, it can be seen that the 3.55 \AA peak is well beyond the limit of error of any d spacing which can belong to the sanbornite structure.

Douglass (1958) further states that: "An uninterpreted powder pattern for $\text{BaO} \cdot 2\text{SiO}_2$ given by Levin and Ugrinic (1953) contains many lines for which there are no corresponding calculated spacings" and concludes that "in the material of Levin and Ugrinic the preponderant phase was not BaSi_2O_5 of sanbornite structure."

In the present study the material used by Levin and Ugrinic was re-examined. Using a modern high angle diffractometer it was found to have an x -ray diffraction pattern essentially the same as that reported in 1953. However when the material was reheated at temperatures below about 1350° C. for many hours it slowly transformed to a low temperature form. This phase has an x -ray powder pattern the same as that reported by Douglass (1958) for sanbornite. Material having the sanbornite structure slowly inverted above about 1350° C. to the high- BaSi_2O_5 form. This form is characterized especially by the diffraction peak at about 3.55 \AA or $25.1^\circ 2\theta$ for $\text{CuK}\alpha$ radiation.

Therefore polymorphism in barium disilicate has been definitely established. The phase transformation takes place at about 1350° C. and is sluggish but reversible. The indices of refraction of the two polymorphs differ by less than two thousandths. Thus the two forms can only be distinguished with confidence by the x -ray diffraction method. The x -ray patterns reported by Austin (1947) and by Levin and Ugrinic (1953) represent the high temperature form of BaSi_2O_5 . The pattern of sanbornite given by Douglass (1958) is the first x -ray pattern of the low temperature form of BaSi_2O_5 to be published.

After the elucidation of the crystal structure of sanbornite, Douglass (1958) also questioned the occurrence of solid solution in BaSi_2O_5 . He doubted that the sanbornite structure could accommodate sufficient change in the BaO to SiO_2 ratio to account for the complete solid solution between $\text{BaO} \cdot 2\text{SiO}_2$ and $2\text{BaO} \cdot 3\text{SiO}_2$, as first reported by Escola (1922). The solid solutions reported to occur at the liquidus are with the high temperature form of BaSi_2O_5 rather than with sanbornite. However, the doubt raised by Douglass' work is probably still valid because of the similarity of the two powder patterns.

A re-examination of the sub-system $\text{BaO} \cdot 2\text{SiO}_2$ - $2\text{BaO} \cdot 3\text{SiO}_2$ has shown that there is actually little or no solid solution in the binary system but that other discrete phases occur. A complete discussion of this sub-system will appear in the Journal of Research of the National Bureau of Standards, **62**, May 1959; RP 2953.

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THE AMERICAN MINERALOGIST, VOL. 44, MARCH-APRIL, 1959

MELTING OF CALCITE IN THE PRESENCE OF WATER

P. J. WYLLIE AND O. F. TUTTLE, *College of Mineral Industries, The Pennsylvania State University*

In a recent note in this journal Paterson (1958) described the partial melting of calcite in the presence of water and carbon dioxide at temperatures "around 900° C. to 1000° C." at a total pressure of 50 bars. The ratio of water vapor to carbon dioxide in the system was not known.

This note is to point out that a program designed to throw light on the origin of carbonatites has been underway for more than two years. The authors are investigating the phase relations in the system CaO - MgO - CO_2 - H_2O as part of the overall program on carbonatites and calcite has

been partially melted at 740°C . in the presence of water vapor at a pressure of 1000 bars. In compositions on the CaO side of the calcite-water join in the system $\text{CaO-CO}_2\text{-H}_2\text{O}$, melting begins at $675 \pm 5^{\circ}\text{C}$. at the same pressure. The joins calcite-water, magnesite-water, and dolomite-water are being studied and, to illustrate the general nature of our results, we will describe briefly the phase relations in a portion of the system $\text{CaO-CO}_2\text{-H}_2\text{O}$ as deduced from runs completed in the join $\text{CaCO}_3\text{-H}_2\text{O}$. These data were presented at the Annual Meeting of the Geological Society of America, St. Louis, Missouri, 1958 (Tuttle and Wyllie, 1958).

Runs were made in small sealed platinum capsules containing measured quantities of water and calcite, the calcite being in the form of a single cleavage rhomb. A preliminary isobaric equilibrium diagram for this join at a pressure of 1000 bars is given in Fig. 1A. The diagram is based on 200 runs determined in the temperature interval from 600°C . to 1200°C . and in the composition range from 10 to 100 weight per cent

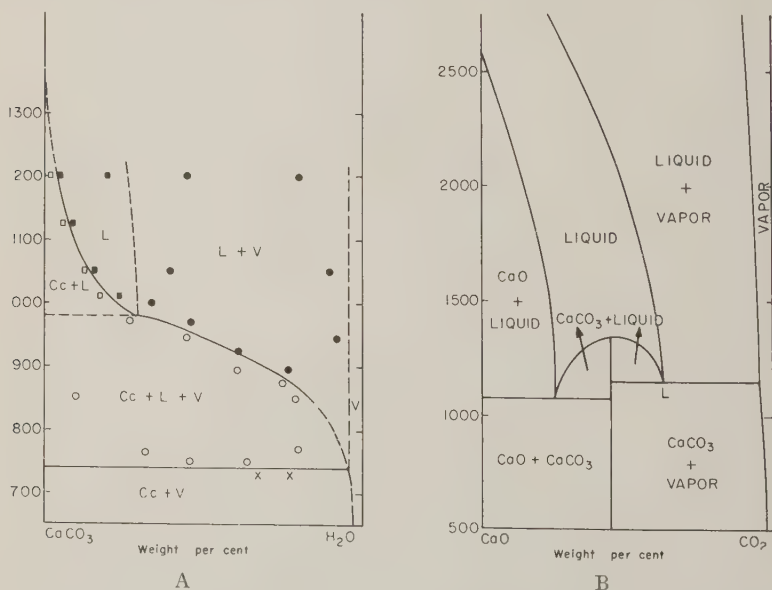


FIG. 1. (A) Isobaric equilibrium diagram for the join calcite-water of the ternary system $\text{CaO-CO}_2\text{-H}_2\text{O}$ at 1000 bars pressure. Cc—calcite; L—liquid; V—vapor. The diagram shows the phase fields intersected by the join at this pressure. When liquid and vapor co-exist their compositions do not lie in the plane of the section (See Fig. 2).

(B) Schematic isobaric equilibrium diagram for the system CaO-CO_2 at 1000 bars pressure. This is believed to represent the probable phase relations, although the positions of the isobaric invariant lines $\text{CaO} + \text{liquid} + \text{CaCO}_3$ and $\text{CaCO}_3 + \text{liquid} + \text{vapor}$ are not yet known.

calcite. The liquid developed along this join quenches to a white compacted powder, which is quite distinct from the transparent cleavage fragment initially sealed into the capsule. The powder consists of a mixture of dendritic calcite and $\text{Ca}(\text{OH})_2$ (portlandite), the proportions of which indicate the composition of the liquid. Much information was obtained from the physical character of the platinum capsules and of the charges within the capsules at the end of each run.

At a pressure of 1000 bars calcite begins to melt at 740°C ; a small amount of calcite dissociates, thus releasing carbon dioxide to the vapor phase, and liquid develops from the CaO , CaCO_3 and water vapor. Only a trace of liquid is developed at this temperature and for charges containing less than 30 weight per cent water, the presence of liquid was not established. Within the three-phase region $\text{CaCO}_3 + \text{liquid} + \text{vapor}$ the proportion of liquid developed at any temperature increases with increasing water content; for a charge of fixed composition the proportion of liquid increases with increasing temperature.

Whenever vapor and liquid co-exist, their compositions do not lie on the join. For this reason it is difficult to discuss the phase relationships without reference to the ternary system $\text{CaO}-\text{CO}_2-\text{H}_2\text{O}$. Vapor is apparently absent between 980°C . and 1200°C . for compositions containing less than 30 weight per cent water, and this part of the join is binary. As shown in Fig. 1A, a one-phase field of liquid is intersected by the join in this region.

As an aid to understanding the experimental results in the ternary system $\text{CaO}-\text{CO}_2-\text{H}_2\text{O}$, a schematic isobaric section through the PTX model of the binary system $\text{CaO}-\text{CO}_2$ is shown in Fig. 1B. These are the probable phase relations at 1000 bars. At this pressure calcite melts congruently and the congruent liquid must be heated well above the melting temperature before a vapor phase appears. It should be noted that two three-phase univariant surfaces are intersected at this pressure, one representing the equilibrium $\text{CaO} + \text{liquid} + \text{CaCO}_3$ and the other $\text{CaCO}_3 + \text{liquid} + \text{vapor}$. The univariant surfaces intersect the isobaric section in straight invariant lines. As water is added to the binary system at constant pressure these invariant three-phase lines move into the isobaric three-dimensional ternary model and generate three-phase spaces which are bounded by three surfaces representing two-phase equilibria. For example, the invariant line $\text{CaCO}_3 + \text{liquid} + \text{vapor}$ becomes a three-phase space bounded by the surfaces $\text{CaCO}_3 + \text{liquid}$, $\text{CaCO}_3 + \text{vapor}$, and $\text{liquid} + \text{vapor}$. The solid phases do not change in composition, but the compositions of the liquid and vapor become ternary. Each of the stability fields shown in the isobaric section of Fig. 1B also extends into the isobaric prism when water is added as a third

component, and each becomes a space separated by the curved surfaces which are generated by the field boundaries of Fig. 1B as they move into the solid model.

It is convenient to visualize the phase relations in the ternary isobaric prism by passing isothermal planes through the model. The isothermal intersection of a three-phase space is a triangle, with the composition of each phase given by the apices, and the intersections of the curved surfaces separating the one and two-phase spaces are curved lines.

Using the information obtained for the join $\text{CaCO}_3\text{-H}_2\text{O}$, it is possible to sketch the isobaric isothermal planes for the system $\text{CaO-CO}_2\text{-H}_2\text{O}$ (Fig. 2). The three-phase triangle calcite+liquid+vapor is bordered by two-phase fields of calcite+vapor, calcite+liquid, and liquid+vapor (Fig. 2B). A one-phase field of liquid, extending from the liquid corner

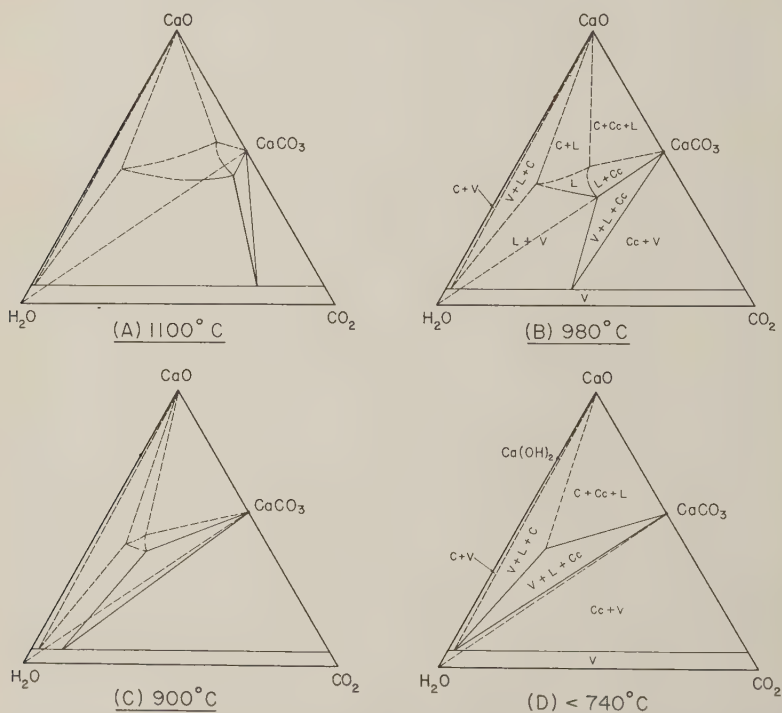


FIG. 2. Schematic isobaric isothermal planes in the system $\text{CaO-CO}_2\text{-H}_2\text{O}$ deduced from data determined for the join calcite-water. C— CaO ; Cc— CaCO_3 ; L—liquid; V—vapor. For compositions on the join calcite-water, melting begins at 740°C . when the calcite-vapor side of the three-phase triangle calcite+liquid+vapor crosses the join. The ternary isobaric invariant temperature illustrated in (D) is $675 \pm 5^\circ\text{C}$.

of the three-phase triangle, intersects the calcite-water join for temperatures above 980°C . At 980°C . (Fig. 2B) the calcite-liquid side of the three-phase triangle crosses the calcite-water join, and at lower temperatures the vapor deficient region of the ternary system is completely within the area $\text{CaO-CaCO}_3\text{-H}_2\text{O}$. The liquid corner of the three-phase triangle is the locus of a liquidus field boundary which passes from a point in the binary system CaO-CO_2 (such as L in Fig. 1B) across the calcite-water join at about 30 weight per cent water at 980°C , (Fig. 2B) to the ternary isobaric invariant point indicated in Figure 2D. Here the last trace of liquid disappears. The temperature of this point is $675 \pm 5^{\circ}\text{C}$. at 1000 bars pressure and $650 \pm 5^{\circ}\text{C}$. at 2000 bars pressure. In addition to the three-phase triangle calcite+liquid+vapor, there must be two other three-phase triangles in the system, $\text{CaO+calcite+liquid}$ and CaO+liquid+vapor , which meet at the ternary isobaric invariant point as shown in Fig. 2D. Below the invariant point the three triangles are replaced by one triangle in which the phases CaO+calcite+vapor are stable. The successive positions of the vapor corners of the three-phase triangles are not known, but with decreasing temperature they must change from pure carbon dioxide to almost pure water. At 1000 bars pressure portlandite becomes stable in the binary system $\text{CaO-H}_2\text{O}$ below 750°C . (Majumdar and Roy, 1956), so there must be a three-phase triangle $\text{CaO+Ca(OH)}_2\text{+vapor}$ in the space shown as CaO+vapor in Fig. 2D. This need not concern us in the present discussion.

The temperature of beginning of melting of calcite for any composition on the join calcite-water is the temperature at which the join intersects the three-phase region calcite+liquid+vapor. This temperature is just above the isobaric ternary invariant point in the system if the relationships are as shown in Fig. 2D; it is believed to be the best interpretation of the available data. The phase relations within the ternary system can be determined more accurately when runs for other joins are completed. The intersection of the join $\text{CaCO}_3\text{-H}_2\text{O}$ with the three-phase triangle $\text{CaCO}_3\text{+liquid+vapor}$ at 900°C . (Fig. 2C) shows how rapidly the proportion of liquid decreases as the percentage of water in the charge decreases. This probably explains why the presence of liquid at 750°C . was not established for charges containing less than 30 weight per cent water. At 750°C . the calcite-vapor side of the three-phase triangle must be almost coincident with the join $\text{CaCO}_3\text{-H}_2\text{O}$.

The vapor phase in Paterson's experiments consisted of a mixture of carbon dioxide and water in unknown proportions. Melting would begin when the three-phase triangle calcite+liquid+vapor crosses the line

joining calcite to the composition of the vapor in the pressure vessel. This temperature would be lowered as the proportion of water in the vapor is increased. The phase relationships in the ternary system corresponding to the beginning of melting in Paterson's experiments could correspond to the condition shown in Figs. 2B or 2C which would represent a temperature well above the minimum liquidus temperature in the ternary system $\text{CaO}-\text{CO}_2-\text{H}_2\text{O}$ at the same pressure. Paterson concluded that calcite, in the presence of water and carbon dioxide at a total pressure of 50 bars, will melt some three to four hundred degrees below the melting point in the presence of carbon dioxide alone. Liquid develops at considerably lower temperatures (650°C.) at 2000 bars pressure within the system $\text{CaO}-\text{CO}_2-\text{H}_2\text{O}$.

The presence in many natural carbonate rocks of additional components such as alkalis would probably lower the melting temperatures further and increase the amount of liquid developed at low temperatures. Some carbonate liquids can be quenched to a glass by extremely rapid cooling, but in nature such rapid cooling is unlikely and evidence of melting in rocks will be difficult to find.

The presence of even a trace of interstitial liquid in a rock will greatly modify its physical character and hasten the attainment of chemical equilibrium. The partial melting of carbonates in the presence of water vapor at temperatures as low as 650°C. and at pressures reached at relatively high levels in the earth's crust has important petrological implications in regard to:

- (1) Emplacement of carbonatites. Many field geologists have long been convinced that carbonatites were emplaced as liquid magmas. Previous experimental data has suggested that the existence of carbonatite liquids in nature was unlikely, but present results prove that carbonatite magmas can exist at moderate temperatures and pressures.

- (2) Deformation of limestones. The development of a small quantity of intergranular liquid in massive limestone would greatly facilitate its deformation.

- (3) Contact metamorphism of limestones. PT conditions at many igneous contacts must be suitable for the development of liquid in limestones. Moreover, there would sometimes be a continuous supply of water and other volatile materials passing from the igneous magma through the aureole; this would produce more melting than could occur in a closed system. The presence of liquid in an impure limestone would aid recrystallization and growth of new minerals.

The carbonate dikes and carbonatite-like bodies associated with the skarn zones of limestone "granite" contacts may also represent intrusion

of carbonate liquid. Burnham (1958) has described dike-like masses of carbonate-microcline-oligoclase-quartz which grade outward from the contact to carbonate veins enclosing euhedral crystals of epidote.

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Note added in press: Study of the system $\text{CaO-CO}_2\text{-H}_2\text{O}$ has now been completed. The general relations deduced above have been confirmed experimentally at 1000 bars pressure, but some modifications are required because it was discovered that portlandite does not dissociate, but melts at 835 C., and calcite melts incongruently, rather than congruently. Detailed results will be published shortly.

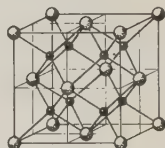
M.S.A. MEMBERSHIP NOTICE

The Council authorized the preparation of an attractive notice to advertise membership in our Society (*Am. Mineral.*, **43**, 1958, 350). The



MINERALOGY

MINERALOGICAL SOCIETY of AMERICA

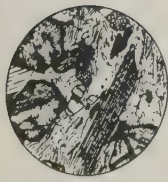


CRYSTALLOGRAPHY

Anyone interested in mineralogy, petrology, crystallography and allied sciences may become a member of the Mineralogical Society of America. The *American Mineralogist*, the bimonthly journal of the Society, which averages about 200 pages each number, is free to members. The annual dues are \$4.00; there is no initiation fee. To obtain a membership application blank, see _____ or

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PETROLOGY



GEOCHEMISTRY

reproduction presented herewith does not do full justice to the 9 by 12 inch form on heavy stock, since it fails to show that the name and seal of our Society are done in red overprint. It is hoped that those in colleges or universities or industrial laboratories or in other callings where a suitable bulletin board is available will drop a postal to the undersigned or the Treasurer requesting a copy for posting. The notice has a blank line on which it is intended will appear the name of a local individual who will be glad to sponsor new members. Membership forms are available from the Treasurer or any member of the membership Committee. One of the better ways to advance the mineralogical sciences is by the wider dissemination of our journal; thus please cooperate.

The teaching profession should be interested in the following quotation paraphrased from a portion of a recent letter from Professor R. A. Hoppin: "It is good to get students interested in the *American Mineralogist*; they too can find much in it that will be useful. Once they get the habit of receiving it, they are very apt to continue their subscription on leaving school. On the other hand once outside the catalytic effect is gone in most cases and they probably won't be interested in starting a new subscription. I know that I felt a certain pride the first time I received a professional journal. I started two publications under student rates. I have continued getting them although I doubt that I would have subscribed to them later at the regular rate."

D. JEROME FISHER, *Chairman*
Membership Committee
(see page 459, this issue)

(Note by editor: Drs. D. J. Fisher and A. A. Levinson are responsible for the design of the membership notice.)

INTERNATIONAL MINERALOGICAL ASSOCIATION

The first general meeting of the IMA since its foundation last year in Madrid, will be held in Zürich, Switzerland, from Monday, August 31 to Thursday, September 3, 1959, and will have a full program. This includes meetings of the Commissions dealing with Abstracts, Mineral Data, Museums, and New Minerals, most of which will be assembled for the first time. The business meeting of delegates (at which representatives from at least 16 nations are expected) will have to deal with certain extensions and modifications of the constitution, prepared by the Executive Committee at a meeting held in Zürich from January 19 to 22, 1959 and attended by five members (Amoros, Fisher, Onorato, Parker, Wickman).

Two symposia will be held during the general meeting. The first of these (September 1 and 2) will deal with *Alpine Fissure Minerals* [Speakers: Epprecht, (Zürich), Fagnani, (Milan), Grigoriev, (Lenin-

grad), Laves, (Zürich), Lietz, (Hamburg), Lovering, (Denver), and Parker, (Zürich)]. The second (September 3, Chairman: Professor M. J. Buerger) will have *Twinning* as its topic [Speakers: Amoros, (Madrid), Buerger, (Cambridge, Mass.), Curien, (Paris), Donnay, (Baltimore), Holser, (La Habra, Cal.), Ito, (Tokyo), and a Russian speaker].

Two concurrent field excursions are planned from Friday, September 4 until Monday, September 7 (inclusive). Of these the "*East*" *Excursion* under the leadership of Professors Staub and Burri and others will visit the Göschener Alps (Granites of the Aar Massif), the mineralized marbles of Castione, the Region of San Bernardino and Maloja in the Engadin, the granite of Val Bregaglia (Bergell) with pegmatites and injection phenomena, and finally the manganese ore deposits of Falotta with characteristic minerals such as tinzenite (Mn-axinite), etc.

The "*West*" *Excursion* to be led by Professor Ernt Niggli (Bern) and others will go from Zürich via Luzern and the Grimsel Pass (granites, gneisses, etc. of the Aar Massif) to Binn (rare sulfosalts in dolomite), thence to Andermatt and Val Maighels (garnetiferous lime-silicate rocks), Sedrun (Alpine fissure mineral deposits), and the Lukmanier Pass (metamorphic rocks and minerals).

Attendance at the symposia will be open to all, as will participation in the excursions, though priority in registering for the latter must be given to the delegates and others attending the General meeting in an official capacity. A more detailed program will be obtainable in the near future from the Secretary of the IMA, Professor J. L. Amoros, Museo de Ciencias Naturales, Castellana 84, Madrid, Spain.

FREQUENCY CONTROL SYMPOSIUM

The 13th annual Frequency Control Symposium, sponsored by the Signal Research and Development Laboratory, will be held May 12-14 at the Berkeley-Carteret Hotel, Asbury Park, New Jersey. A total of 62 technical papers will be presented.

The sessions will be concerned with the fundamental properties of quartz, piezoelectric resonators, crystal measurements, stable oscillator circuits, ultra precision crystals, crystal standards, frequency synthesis and measurements, precision filters, atomic and molecular beam devices, and gas cells.

CONFERENCE ON APPLICATIONS OF X-RAY ANALYSIS

The eighth annual conference on Applications of X-Ray Analysis, sponsored by the Metallurgy Division of the Denver Research Institute of the University of Denver, will be held August 12-14 at the Stanley Hotel, Estes Park, Colorado. The deadline for titles and author information is May 1, 1959. Conference Chairman: William M. Mueller, Denver Research Institute, University of Denver, Denver 10, Colorado.

ERRATUM

In the January-February, 1959 issue of *The American Mineralogist*, page 5, the d spacing for $hkl=411$ (Table 1) should read 1.299 instead of 1.229.

BOOK REVIEWS

HANDBOOK OF CHEMICAL MICROSCOPY, Volume 1. Principles and Use of Microscopes and Accessories. Physical Methods for the Study of Chemical Problems. Third Edition, by ÉMILE MONNIN CHAMOT and CLYDE WALTER MASON. John Wiley and Sons, Inc., New York, N. Y., 1958. xii+502 pages, 1 fold-in color plate Michel-Levy scale of birefringence, 125 Figures, Price, \$14.00.

Mineralogists, particularly those engaged in the study of chemicals, ceramic and similar products, will welcome the Third Edition of this well-known and most useful handbook. From a practical point of view, this book has two important and distinguishing features: (1) It contains many suggestions and remedies for the everyday as well as the unusual problems which face the practicing microscopist, and (2) It has several thousand literature references, many to special problems and applications, all arranged in convenient footnote form.

Several of the chapters of the Third Edition under review have been revised and modernized to a great extent from the previous edition (published in 1938); but, in some chapters, very few changes have been made. This edition contains a short chapter on the electron microscope. The discussion of the relation of optical properties to structure of crystals and aggregates has been enlarged and given chapter status; the added material is mostly basic crystallography. Newer techniques, such as phase contrast microscopy, are conveniently added to previously existing chapters.

In any attempt to completely modernize a textbook some of the older material might be expected to survive and we have no exception here. The reader will find several illustrations of somewhat antiquated equipment. In general, the illustrations are well chosen but this reviewer believes the petrographic microscope deserves more than a three inch sketch (of an older model) especially when, for example, a photographic eyepiece and attached camera receive about one-half page. Some examples of past eras are still to be found within the text; for example, we might infer (p. 275) that the "cap" analyzer is still rather commonly used. These points are minor especially when we consider that literature references as late as 1958 are included.

The entire subject of optical crystallography is covered in about 50 pages, distributed among several chapters, in an order which would be unappealing to most mineralogists. Although all important subjects are covered, many are quite sketchy. The authors, however, do not intend their coverage to be adequate for all workers in the field and refer the reader to standard texts, many of them mineralogical. A new feature of the Third Edition is the Michel-Levy scale of birefringence provided by the Bausch and Lomb Optical Company and adapted from "Optical Mineralogy" by Rogers and Kerr. Unfortunately, the mineral names on the chart by Rogers and Kerr have been removed resulting in a chart with neither minerals nor chemical compounds illustrating the various birefringences. Under any condition, the usefulness of a Michel-Levy birefringence chart in this book is questionable.

There can be no doubt, however, that this book will be, quite deservedly, well received and continue to be one of the most widely used on the subject. In the opinion of the reviewer, this book illustrates the fact that mineralogists, with their superior training in petrographic methods, are far better equipped to tackle many problems in the general areas of chemical microscopy than are microscopists entering the field from other sciences. Teachers of optical mineralogy would do well to impress upon their students the value of the polarizing microscope in the ever increasingly important study of textiles, polymers, pigments, and other organic and inorganic materials of interest to industry.

A. A. LEVINSON
Dow Chemical Company
Freeport, Texas

GRUNDZÜGE DER LITHOLOGIE; Lehre von den Sedimentgesteinen. By L. B. RUCHIN, edited in the German edition by A. Schuller; translated from the Russian by J. Bar-nitzke. 806 pages, 304 illustrations, 46 tables. Akademie-Verlag, Mohrenstrasse 39, Berlin W 8. 56 DM. 1958.

In translating this work from Russian to German, the editor and translator have made a valuable work more readily available to European and American geologists and in the course of this service have completed an outstanding job of organizing and presenting a vast amount of information on sedimentary rocks. Reportedly French and Chinese translations of this work also have been prepared. For 25 years the author has been associated with the Geological Faculty of Leningrad University, at which he presently is Professor. The book has been awarded the State Prize of the USSR.

Fundamentally the book deals with the development of sedimentary regions and their relationships to the preceding tectonic history, with special reference to erosional and depositional features. In the developing of this theme, the author has organized his material in five parts: I. Sedimentary rocks; II. Stages and conditions for the formation of sedimentary rocks; III. Facies and methods of facies analysis; IV. Sedimentary formations, and V. The present geologic epoch and its sedimentary deposition. Probably the only United States book of comparable plan and scope is "Stratigraphy and Sedimentation" by Krumbein and Sloss (W. H. Freeman and Co.).

Topics discussed in Part I include: lithology as a science, divisions of sedimentary rocks, clastic rocks, argillaceous rocks, chemical and organic sedimentary rocks, and sedimentary rocks as geologic units. In Part II are included discussions of weathering, transportation, deposition, consolidation, and the most important theories on the formation of sedimentary rocks. The section on facies is divided into chapters on general characteristics, and field and laboratory methods of facies analysis. Part IV includes descriptions of formations and their distribution, the most important geosynclinal formations, tectonically affected formations in transitional regions, tectonically affected formations on shelves, some laws in the formation of sedimentary formations. The last part deals with the topics, recent ocean sediments, recent deposits in marine geosynclinal basins, recent deposits on submarine shelves, recent lagoonal deposits, recent continental deposits in low relief regions under a moist temperate climate, recent continental deposits in low relief regions under a hot arid climate, recent continental deposits in mountains, and general characteristics of the present geologic epoch and its deposits. These subject titles (28 chapters) serve as a sample of the coverage of the work.

Charts, diagrams, and line drawings are adequate to good throughout, but reproductions of photographs are generally drab and detail-poor. The list of references includes 161 entries, given first in Russian, followed by the German translation. A very high percentage of the references is to Russian publications. There are two indexes—author and subject.

The American student of sedimentary rocks and processes will look in vain for reference to some of the familiar publications in English and to classic examples with which he is familiar. It is in this substitution of European (and particularly of Russian) ideas, examples, and references that the monograph will be of especial value to all serious workers in sedimentary petrogenesis in this country. To the relatively scanty library of general works in this field, its availability is welcome and its inclusion is desirable.

E. WM. HEINRICH
The University of Michigan

NEW MINERAL NAMES

Uramphite

Z. A. NEKRASOVA. A hydrous uranyl ammonium phosphate (uramphite), $\text{NH}_4(\text{UO}_2)(\text{PO}_4) \cdot 3\text{H}_2\text{O}$. *Voprosy geologii urana, Atomizdat*, 1957, p. 67-72; from a translation by Consultants Bureau 1958, p. 56-60.

A microchemical analysis by R. P. Khodzhaevaya gave UO_3 68.70, P_2O_5 15.63, NH_4 4.60, H_2O 11.00, sum 99.93%, giving the ratio $\text{P}_2\text{O}_5:\text{UO}_3:\text{NH}_4:\text{H}_2\text{O}=1:2.2:2.3:5.5$, or $\text{NH}_4(\text{UO}_2)(\text{PO}_4) \cdot 3\text{H}_2\text{O}$. The mineral is easily soluble in weak cold HCl and in weak warm HNO_3 . D.T.A. study shows endothermal breaks at 150° (loss of H_2O) and 500° (loss of NH_3 and destruction of crystal lattice), and an exothermic break, at 630° (crystallization of uranium pyrophosphate). Fluoresces medium yellow-green; the intensity of fluorescence decreases as the mineral is heated and vanishes when the mineral is heated at 500° for a long time.

Color bottle-green to pale green, luster vitreous. G. 3.7. Cleavage rather distinct in two directions. Optically negative, $2V$ $0-3^\circ$, α 1.564, $\beta=\gamma$ 1.585. Pleochroism X colorless, Y and Z pale green.

X-ray powder data are given for uramphite and for the synthetic material (method of synthesis not stated); 77 lines are given. The strongest lines are 3.78 (10), 2.22 (9), 1.694 (9-8), 2.75 (8), 2.12 (8), 3.24 (7), 1.399 (7), 1.267 (7).

The mineral occurs in the oxidized zone of a uranium-coal deposit in fractures in the coal 20-50 m. below the surface. The locality is not stated, as usual. The mineral occurs in square tablets (up to 0.2×0.2 mm.), in small rosettes, and as lichen-like deposits.

The name is for the composition.

DISCUSSION.—Perhaps a member of the meta-autunite or meta-torbernite groups.

MICHAEL FLEISCHER

Ursilite (Calcium ursilite, Magnesium ursilite)

A. A. CHERNIKOV, O. V. KRUMETSKAYA, and V. D. SIDEL'NIKOVA. Ursilite—a new silicate of uranium. *Voprosy geologii urana*, 1957, p. 73-77; from a translation by Consultants Bureau 1958, p. 61-65.

Analyses (by O.V.K. and V.D.S.) are given of four samples.

	1	2	3	4
SiO_2	26.90	26.67	26.92	26.80
Al_2O_3	0.24	—	—	—
Fe_2O_3	0.13	0.27	0.14	—
CaO	5.75	6.67	6.31	0.79
MgO	0.95	1.55	2.49	6.72
K_2O	none	—	none	none
Na_2O	none	—	none	none
UO_3	50.57	49.80	50.40	50.32
H_2O^-	8.83	8.3	—	—
H_2O^+	6.90	5.83	14.7	14.9
Total	100.27 ^a	99.09 ^b	100.96	99.53

^a Given as 100.33.

^b Given as 99.12.

These give for the ratios

	(CaO+MgO)	UO ₃	SiO ₂	Total H ₂ O
(1)	0.72	1.00	2.55	5
(2)	0.9	1.00	2.55	4.5
(3)	1.0	1.00	2.53	4.7
(4)	1.08	1.00	2.55	5.7

These correspond to $2(\text{Ca, Mg}) \cdot 0.2\text{UO}_3 \cdot 5\text{SiO}_2 \cdot 9\text{--}10\text{H}_2\text{O}$, with Ca predominant in Nos. 1–3 (calcium ursilite), Mg in No. 4 (magnesium ursilite). The mineral is easily soluble in acids, especially when warm, and yields silica gel. It turns black when heated.

The mineral is lemon-yellow and fluoresces greenish-yellow under ultra-violet light. Brittle, hardness 3. G. (pycnometer) 3.034 (calcium ursilite), 3.254 (magnesium ursilite). Optically biaxial, negative, indices α 1.548, γ 1.556 (calcium ursilite), α 1.543, γ 1.550 (magnesium ursilite). Extinction parallel, elongation positive. Weakly pleochroic with X pale green, Z green.

X-ray powder data are given for the material of analyses no. 1 (35 lines) and no. 4 (10 lines); these agree closely. The strongest lines are for calcium ursilite: 3.37 (10), 3.02 (10), 4.56 (9), 1.828 (9), 1.158 (9), 1.120 (9), 5.04 (8), 1.562 (8), 1.378 (8), 1.345 (8), 1.075 (8), 1.054 (8), 2.63 (7), 2.51 (7), 2.22 (7), 1.038 (7); for magnesium ursilite 4.98 (10), 3.06 (10), 2.30 (9), 2.07 (9), 4.58 (8), 3.37 (8), 2.42 (6), 1.981 (6), 1.798 (6).

The mineral occurs as earthy or nodular incrustations, rarely in radiating spherulites, along joints in quartz porphyries, associated with kaolinite and calcite, and rarely with uranophane, sklodowskite, and kasolite. The locality is not given.

The name is for the composition.

DISCUSSION.—The optical properties are very different from those of the calcium uranyl silicates uranophane and beta-uranophane and the magnesium uranyl silicate sklodowskite. The nearly identical optics and x-ray powder patterns of the Ca and Mg minerals are surprising, and further study, especially of crystal structure, is highly desirable.

M. F.

Woodfordite

JOSEPH MURDOCH AND ROBERT A. CHALMERS. Woodfordite, a new mineral from Crestmore, California. *Bull. Geol. Soc. Am.*, **69**, p. 1620–1621 (1958) (abs.).

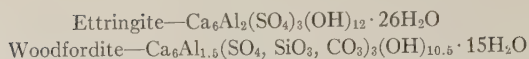
Woodfordite occurs at the 910 ft. level of the Commercial Quarry, Crestmore, California, associated with afwillite and calcite in veins cutting massive contact rock. Analysis (not given) leads to approximately $2\text{Ca}[(\text{SiO}_3)_{0.15}(\text{CO}_2)_{0.2}(\text{SO}_4)_{0.65}] \cdot 2\text{Ca}(\text{OH})_2 \cdot \text{Al}(\text{OH})_3 \cdot 10\text{H}_2\text{O}$ (CO_2 error for CO_3 ? M.F.). Readily soluble in HCl, usually with slight effervescence. When heated, loses water readily, turns white and fibrous, but apparently does not fuse.

Transparent, colorless, hardness 2.5, G. 1.85. Optically uniaxial, negative, ω 1.465, ϵ 1.455.

Hexagonal, in prisms, with $(10\bar{1}0)$ dominant, (0001) normally present, $(10\bar{1}2)$ sometimes present, (7078) (?) usual. Cleavage prismatic perfect. X-ray study shows $P6_3$ or $P6_3/m$, a_0 22.32, c_0 21.33 Å. The strongest lines are in Å., 9.67–10, 5.6–9, 3.88–6, 2.568–6, 2.204–6.

The name is for Professor A. O. Woodford of Pomona College who found the mineral.

DISCUSSION.—This is close in composition and physical properties to ettringite. The compositions may be compared:



M. F.

Heinrichite, Meta-heinrichite
Sandbergerite, Meta-sandbergerite

E. B. GROSS, A. S. COREY, R. S. MITCHELL, AND KURT WALENTA. Heinrichite and meta-heinrichite, hydrated barium uranyl arsenate minerals. *Am. Mineral.*, **43**, 1134–1143 (1958).

Arsenuranocircite

L. N. BELOVA. Arsenuranocircite, $\text{Ba}(\text{UO}_2)_2(\text{AsO}_4)_2 \cdot 8\text{H}_2\text{O}$. *Proc. 2nd Geneva Conf. on Peaceful Uses of Atomic Energy*, **1958**, vol. 2, p. 294.

The data from the Russian paper are as follows: Analysis by Y. S. Nesterova and L. E. Novorossova gave BaO (given as Ba, but the mol. ratio indicates this to be BaO M.F.) 12.99, As_2O_5 23.46, UO_3 50.03, H_2O 12.00, sum 98.48%. Occurs in a pitchblende-molybdenite deposit as large tabular light green plates associated with metazeunerite and barium uranophane. Brightly fluorescent, like uranocircite. Optically uniaxial, negative, ω 1.632, ϵ 1.623. Unindexed x-ray powder data are given; (43 lines); the strongest are 9.82 (8), 3.68 (8), 1.461 (8), 8.64 (5), 4.04 (5), 2.20 (5), 1.656 (5).

DISCUSSION.—The identity of “arsenuranocircite” with meta-heinrichite is evident, despite the discrepancies in birefringence and x-ray spacings. The use of three different names appearing almost simultaneously is unfortunate; it is not clear which has priority. The names heinrichite and meta-heinrichite first appeared in a note by Walenta in *Der Aufschluss*, **9**, No. 10, p. 247–248, October, 1958. The name sandbergerite has been dropped by mutual agreement, which leaves a choice between heinrichite and arsenuranocircite (the mineral described should have been called meta arsenuranocircite). The description of heinrichite is much more complete.

M. F.

Meta-Kirchheimerite

K. WALENTA. Die sekundären Uranmineralien des Schwarzwaldes. Preliminary report, *Techn. Hochschule, Stuttgart*, **1958**, from an abstract in *Bull. soc. franc. mineral. et. crist.*, **81**, p. 67–68 (1958).

A member of the meta-torbernite group, probably $\text{Co}(\text{UO}_2)_2(\text{AsO}_4)_2 \cdot 8\text{H}_2\text{O}$. Microchemical analysis showed major Co, U, and As, a little Fe, and traces of Ni. In pale rose crusts and tabular crystals. Cleavage (001) excellent. Luster pearly on the cleavage. Hardness 2–2½. G. above 3.33. Optically uniaxial to biaxial, negative, $2V$ 0–20°. Color greenish-yellow to colorless, not pleochroic, ω 1.644, ϵ 1.617 (both ± 0.002).

The strongest x-ray lines are 8.55 (10), 3.56 (10), 5.07 (6), 4.30 (6), 3.00 (6), 3.41 (5), 2.52 (5).

Occurs in the Sophia shaft at Wittichen, Baden, on pitchblende. It is associated with meta-kahlerite, novacekite, meta-heinrichite, and erythrite.

The name is for F. Kirchheimer, director of the Geologisches Landesamt für Württemberg-Baden.

M. F.

Barium uranophane

L. N. BELOVA. Barium uranophane. *Proc. 2nd Internatl. Conf. Peaceful Uses of Atomic Energy, Geneva*, **1958**, v. 2, p. 295.

The mineral occurs in crusts and thin crystals associated with arsenuranocircite and

relicts of pitchblende, soddyite, uranophane, and metazeunerite near a large fault filled by thick barite veins. The crystals have pyramidal terminations. They are bright yellow; fluorescence in ultra-violet green-yellow, a little more intense than normal uranophane.

There was not enough material for chemical analysis. Semiquantitative spectrographic analysis by L. N. Indichenko showed Ba, U, Si-n%, Mo, Al, Ca-0.n%, Mg, Zn-0.0n%, Ti-0.00n%.

The mineral is biaxial, positive, α 1.797, β 1.818, γ 1.820, extinction $c:\gamma = 10-14^\circ$.

X-ray powder data (19 lines) are given; the strongest lines are 7.93 (8), 3.99 (6), 2.14 (5), 1.997 (6), 3.58 (4), 3.38 (4), 2.88 (4), 2.23 (4). "The x-ray data are very similar to those of other uranium silicates, especially of cuprosklodowskite."

DISCUSSION.—Needs further study.

M. F.

Sinicite

HO CHEN-TSI AND CHUN CHI-CHEN. Sinicite, a new mineral, a uranium-bearing titanate. *Kasyua Tunbao* ("Scientia") 1957, No. 12, p. 378; from an abstract by E. M. Bonshedt-Kupletskaya in *Zapiski Vses. Mineral. Obshch.*, 87, No. 4, p. 479 (1958).

The mineral is blackish-brown to reddish-brown, streak reddish-brown, luster resinous. Hardness about 6, very brittle, fracture conchoidal. G. 4.919–4.976.

Analyses gave:

	Blackish-brown	Reddish-brown		Blackish-brown	Reddish-brown
MgO	0.33	0.55	ThO ₂	8.29	7.93
CaO	1.20	1.36	UO ₃	5.98	6.83
PbO	0.84	—	TiO ₂	24.96	
MnO	0.15	0.12	Nb ₂ O ₅	20.85	47.48
FeO	none	—	Ta ₂ O ₅	2.75	
Fe ₂ O ₃	4.44	3.62	H ₂ O ⁻	0.21	0.85
Al ₂ O ₃	1.81	2.08	H ₂ O ⁺	2.53	3.37
Ce ₂ O ₃	7.79	5.74	SiO ₂	0.59	0.31
[Ce] ₂ O ₃	10.87	11.40			
[Y] ₂ O ₃	6.91	7.27		100.50	98.91

The formula is given as



The material, heated at 1000° for 2 hours, gave an x-ray pattern with lines (in Å.): 2.94 (10), 1.63 (8), 1.58 (6), 3.04 (4), 2.56, 2.45, 2.04, 1.92, 1.81 (2), 1.74 (1).

Occurs in a granite pegmatite in China in the zone of coarse feldspar and mica, associated also with quartz, biotite, muscovite, topaz, fluorite, tourmaline, and beryl.

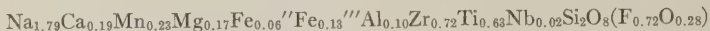
DISCUSSION.—Comparison with the many other minerals of similar composition is needed.

M. F.

Seidozerite

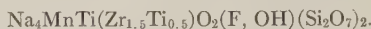
E. I. SEMENOV, M. E. KAZAKOVA, AND V. I. SIMONOV. A new zirconium mineral, seidozerite, and other minerals of the wöhlerite group in alkaline pegmatites. *Zapiski Vses. Mineralog. Obshch.*, 87, No. 5, p. 590–597 (1958) (in Russian).

Analysis by M. E. K. gave SiO_2 31.40, ZrO_2 23.14, TiO_2 13.16, Nb_2O_5 0.60, Al_2O_3 1.38, Fe_2O_3 2.85, FeO 1.06, MnO 4.22, MgO 1.79, CaO 2.80, Na_2O 14.55, H_2O 0.60, F 3.56, sum 101.11%—($\text{O}=\text{F}_2$) 1.49=99.62%. This corresponds to



(Note—should be $(\text{OH})_{0.28}$ M.F.) or approximately $\text{Na}_2\text{Mn}_{0.5}(\text{Zr}, \text{Ti})_{1.5}\text{Si}_2\text{O}_8\text{F}$. This is compared to the formulas of other members of the wöhlerite group; it is close to titanolavenite, but has much less calcium. X-ray spectrographic analysis by I. D. Shevaleevskii gave ZrO_2 23, HfO_2 0.4%; spectrographic analysis showed traces of Pb, Sn, Cr, Be, and Ga. Dissolves with difficulty in HCl. Fuses easily before the blowpipe.

An x-ray study by V. I. Simonov and N. V. Belov, *Doklady Akad. Nauk S.S.S.R.* **122**, 473–476 (1958) suggests that the formula should be written as



Seidozerite is monoclinic, occurring in radiating crystals up to 5×1 cm. in size. Monoclinic, forms noted $c(001)$, $a(100)$, $e(203)$, $b(010)$, $d(011)$, $p(111)$; goniometric data gave $a:b:c=0.780:1.00:2.601$, $\beta 102^\circ 43'$. Rotation diagrams gave space group C_s^2-Pc or C_{2h}^4-P2/c , (test for piezo-electricity negative), $a_0 5.53 \pm 0.03$, $b_0 7.10 \pm 0.04$, $c_0 18.30 \pm 0.1\text{Å}$, $\beta 102^\circ 43'$, $a:b:c=0.779:1.00:2.58$, $Z=4$. (Note—the orientation differs from that generally given for members of this group: $a_0=\frac{1}{2}a$, $b_0=c$, $c_0=2\frac{1}{2}b$. M.F.). Cleavage perfect (001). X-ray powder data by N. N. Sludsko are given; the strongest lines are 2.89 (10), 2.83 (9), 3.21 (7), 3.97 (5), 2.00 (5), 1.792 (5).

The mineral is brownish-red to reddish-yellow in small fragments, translucent with red color. Luster strong, vitreous. Brittle. Hardness 4–5, G. 3.472, 3.87 (calcd. from x-ray data). Optically biaxial, positive, with x 1.725, β 1.758, γ 1.830, $2V$ 68° , plane of optic axes parallel (001), $X=b$, $Z=a$, $Y:c=13^\circ$. Strong dispersion of optic axes, $r>v$. Pleochroism intense, X dark red, Y red, Z light yellow; absorption $X>Y>Z$.

The mineral has been found in four nepheline syenite pegmatites of the Muruai and Uel'-kuai Rivers, Seidozero region, Lovozero massif, Kola Peninsula, where it occurs with microcline, aegirine, nepheline, apatite, pyrochlore, magnetite, ilmenite, titanolavenite, and eudialyte. It is replaced by minerals of the catapleiite type.

The name is for the region.

DISCUSSION.—A member of the wöhlerite group. The entire group needs a thorough x-ray study.

M. F.

Unnamed

L. S. RUPNITSKAYA. Calcium uranium molybdate, $\text{Ca}(\text{UO}_2)_3(\text{MoO}_4)_3(\text{OH})_2 \cdot 8\text{H}_2\text{O}$ *Proc. 2nd Internat. Conf. Peaceful Uses of Atomic Energy, Geneva*, **1958**, **2**, 286.

The mineral occurs as elongated prismatic crystals forming sheaf-like radiating aggregates up to 1.5 mm. long. It has a rich yellow color, with a honey tint. Fluorescence yellow-green, a bit brighter than that of uranophane. ns above 1.758; pleochroic with Z bright yellow, X nearly colorless. Microchemical analysis gave UO_3 61.26, MoO_3 19.90, CaO 4.63, SiO_2 3.86, H_2O 10.90, sum 100.55%; corrected after deducting uranophane, UO_3 58.65, MoO_3 27.20, CaO 3.91, H_2O 10.21, sum 99.97%. Spectrographic analysis showed also 0.01–0.1% Fe and As and traces of Al. Readily attacked by dilute HCl. and HNO_3 .

X-ray powder data (16 lines) are given; the strongest are 7.85 (10), 3.21 (8), 3.89 (6), 8.34 (5), 1.99 (5), 1.855 (4).

The mineral occurs in the lower part of the oxidation zone of hydrothermal uranium-molybdenum veinlets (locality not given).

M. F.

Chalcopentlandite

HANS PAULY. Igdlukunguaq nickeliferous pyrrhotite, texture and composition. A contribution to the genesis of the ore type. *Meddelelser om Gronland*, **157**, No. 3, 1-167 (1958).

The name chalcopentlandite is given to an assumed high-temperature phase, now represented by a two phase aggregate of chalcopyrite (perhaps 10%) in pentlandite, and believed to have been formed by exsolution. Some areas were found by microscopic study (no x-ray data!) that might be "untransformed chalcopentlandite."

M. F.

Simplotite

M. E. THOMPSON, C. H. ROACH, AND ROBERT MEYROWITZ. *Am. Mineral.*, **43**, 16-24 (1958).

Santafeite

MING-SHAN SUN AND R. H. WEBER. *Am. Mineral.*, **43**, 677-687 (1958).

Sherwoodite

M. E. THOMPSON, C. H. ROACH, AND ROBERT MEYROWITZ. *Am. Mineral.*, **43**, 749-755 (1958).

Eskolaite

OLAVI KUOVO AND YRJO VUORELAINEN. *Am. Mineral.*, **43**, 1098-1106 (1958).

Ajoite

W. T. SCHALLER AND A. C. VLISIDIS. *Am. Mineral.*, **43**, 1107-1111 (1958).

Chavesite

JOSEPH MURDOCH. *Am. Mineral.*, **43**, 1148-1156 (1958).

NEW DATA**Messelite, Neomesselite**

F. ČECH AND K. PADERA. Messelit aus den Phosphatnestern im Granit bei Pribyslavice (Bohmen) und das Messelitproblem. *Chem. der Erde*, **19**, No. 4, 436-449 (1958).

Messelite was described from Messel, Hesse, by Muthmann in 1890 as $\text{Ca}_2(\text{Fe}, \text{Mg})(\text{PO}_4)_2 \cdot 2\frac{1}{2}\text{H}_2\text{O}$. In 1940, Wolfe (*Am. Mineral.*, **25**, 738-753, 787-809 (1940)) re-examined type material and showed that it was a mixture of two minerals, namely anapaite plus a mineral he thought to be collinsite. Frondel (*Am. Mineral.*, **40**, 828-833 (1955)) found a mineral at Palermo, N. H., that corresponded in composition with messelite and gave it the name neomesselite.

Restudy of material from the type locality by optical and x-ray methods now shows that the major constituent is identical with the neomesselite of Frondel; this mineral is rimmed by anapaite. Data are also given for samples from Czechoslovakia and Kazakhstan.

DISCUSSION.—Which name to keep is a problem; the authors, in agreement with Professor Strunz, consider that messelite should be retained, since this was the major constituent of the mixture probably studied by Muthmann. I agree; the name neomesselite should be dropped.

M. F.

Taramellite

FLORENZO MAZZI. Riesame della taramellite. *Atti soc. tosc. sci. nat., ser. A*, v. **64**, p. 237-245 (1957).

Re-analysis of taramellite (Tacconi, 1908) showed that the presence of Ti had been overlooked. A new analysis on material dried at 110° gave SiO_2 33.9, TiO_2 7.7, Fe_2O_3 12.2, FeO 3.7, MgO 0.8, BaO 37.5, CaO 1.1, Na_2O 0.8, K_2O 0.1, H_2O^+ 2.1, sum 99.9%. This corresponds to $(\text{Ba}, \text{Ca}, \text{Na})_4(\text{Fe}'', \text{Mg})\text{Fe}_2'''\text{Ti}(\text{Si}_4\text{O}_{12})(\text{OH})_4$, with a little Ti replacing Si. X-ray study showed the mineral to be orthorhombic with a_0 7.05, b_0 13.95, c_0 12.01 Å (all $\pm 0.5\%$), $Z=2$. Space group $C_{2v}^7(\text{Pnm}2_1$ or $\text{Pn}2_1m$) or $D_{2h}^{13}(\text{Pnmm})$. The mineral is apparently to be classed as a sorosilicate.

M. F.

Bismutoferrite, Chapmanite

CHARLES MILTON, J. M. AXELROD, AND BLANCHE INGRAM. *Am. Mineral.*, **43**, 656–670 (1958).

Roscherite

M. L. LINDBERG. *Am. Mineral.*, **43**, 824–838 (1958).

DISCREDITED MINERALS

Truscottite (=Reyerite)

H. STRUNZ, AND H. MICHEELSEN. Calcium-Phyllosilikate. *Naturwissenschaften*, **21**, 515–516 (1958).

In this preliminary note, it is stated that chemical, optical, and x-ray powder data show the identity of truscottite (Hovig, 1914) with reyerite (Cornu and Himmelbauer, 1907). Reyrite has space group $C_{3i}^1-C\bar{3}$, a_0 9.73, c_0 18.72 Å, idealized formula $\text{Ca}_9(\text{OH})_6\text{Si}_{13}\text{O}_{36}$, $Z=2$. Mackay and Taylor, *Mineralog. Mag.*, **30**, 450–457 (1954) found for truscottite a 9.72, c 18.71 Å, unit cell contents near $\text{Ca}_{12}(\text{OH})_{12}\text{Si}_{24}\text{O}_{54}$.

M. F.

Centrallasite (=Gyrolite)

H. STRUNZ, AND H. MICHEELSEN. Calcium-Phyllosilikate. *Naturwissenschaften*, **21**, 515–516 (1958).

Centrallasite (How, 1859) is stated to be identical with gyrolite (Anderson, 1851), as previously reported by Bannister (private communication to M. H. Hey, Chemical Index of Minerals). Gyrolite is hexagonal with a_0 9.80, c_0 22.08 Å. Mackay and Taylor, *Mineralog. Mag.*, **30**, 80–91 (1953) found it to be trigonal with a_0 9.72, c_0 6×22.13 Å. It is suggested that perhaps two modifications exist.

M. F.

Radiophyllite (=Zeophyllite)

H. STRUNZ, AND H. MICHEELSEN. Calcium-Phyllosilikate. *Naturwissenschaften*, **21**, 515–516 (1958).

It is stated that chemical, optical, and x-ray study show radiophyllite (Brauns and Brauns, 1924) to be identical with zeophyllite (Pelikan, 1902).

M. F.

Hypochlorite (Bismutoferrite + Quartz)

CHARLES MILTON, J. M. AXELROD, AND BLANCHE INGRAM. *Am. Mineral.*, **43**, 656–670 (1958).

M. F.

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